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POTASSIUM-ARGON DATING OF SEDIMENTARY ROCKS

BY JOSEPH LIPSON

ABSTRACT

Results in the potassium-argon dating program at Berkeley are reported. Geologically well-classified authigenic sediments ranging from Miocene (12 m.y.) to Givetian (285 m.y.) have been analyzed utilizing equipment designed to date materials of low radiogenic argon content. Age determinations of seven glauconites from the Oligocene (30 m.y.), Eocene (45 m.y.), and Paleocene (55 m.y.) of New Zealand are consistent with the stratigraphy and correspond to the few previous U-Pb and K-A dates for that segment of the geological time scale. Ages of three Miocene glauconite samples from New Zealand are anomalously high compared to those of the Oligocene samples, which may be too low. The age for an Albian glauconite from Canada is too high compared to that of other samples dated. Dating of a Cenomanian feldspar and a Givetian sylvite from Canada is consistent with the Holmes B time scale. Possible explanations are offered for the discrepancies, and further experiments are suggested.

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INTRODUCTION

The potassium-argon dating method utilizes the radioactive decay of K^{40} to A^{40} which results in the accumulation of radiogenic A^{40} in all solid materials containing potassium. Two nuclear constants govern the rate of A^{40} accumulation. The first of these is λ, the total decay constant, which determines the rate at which K^{40} disappears. The second is the branching ratio, R, which is the ratio λ_K/λ_β. The decay constant λ_K is for the decay of K^{40} to A^{40} by K capture, and λ_β is the decay constant for the decay of K^{40} to Ca^{40} by electron emission. The branching ratio, R, is therefore the ratio between A^{40} produced and Ca^{40} produced. If the branching ratio, R, and the total decay constant, λ, are known the age of a mineral can be calculated by determining the content of radiogenic argon and the amount of potassium

in the mineral. The latter measurement immediately gives the amount of K^{40}, since the isotopic composition of potassium is known and does not vary with origin or location (Rankama, 1954, p. 307). The age calculated for a mineral is the length of time necessary to generate the radiogenic argon now found in the mineral.

The potassium-argon method has several advantages. The 1.2-billion-year half life is ideal for dating almost the entire range of geologic time; the almost-universal presence of potassium makes the method applicable in principle to almost any geologic setting; and finally, because of the weak nature of the disintegrations, radiation damage of the host mineral should not result in accelerated argon leakage. The principal difficulties arise from the fact that 99.6 per cent of atmospheric argon is A^{40}. This means that the small quantities of A^{36} and A^{38} (.4 per cent in atmospheric argon) must be used to determine how much of the A^{40} is atmospheric argon. If a quantity of almost pure A^{38} is added to the argon from the mineral, the known quantity of A^{38} serves as an isotopic tracer which makes possible a quantitative determination of the amount of A^{40} in a sample of the argon from a mineral. Because of the addition of A^{38}, the abundance of A^{36} is the only remaining index to the amount of atmospheric argon.

The assumptions are: (1) that the argon in the specimen is a mixture of pure radiogenic argon produced since the mineral was formed and of argon of the same isotopic constitution as present atmospheric argon; (2) that the time of formation of the mineral is short compared to the age of the solidified mineral; and (3) that no argon has escaped from the mineral since its formation. Any departure from the assumptions requires correction of the age determination. If the departures from the assumptions are large and random, the value of the dating method is seriously diminished. If restricted classes of geologic materials deviate from the ideal in a systematic way, however, the method can still be expected to supply useful information.

The history of potassium-argon dating falls into four categories. These are: (1) the general progress in potassium-argon age determination; (2) improvements in the methods of argon extraction; (3) physical measurements on the K^{40}-decay process; and (4) the relatively recent application of the method to sedimentary processes.

Aldrich and Nier (1948, p. 876) first proved that old minerals contain radiogenic argon. They extracted argon from minerals of known age and analyzed them mass spectrometrically. Using the A^{40}/A^{36} ratio as a guide, they found that all the minerals examined contained radiogenic argon and that the amount of radiogenic argon increased with age. Smits and Gentner (1950, p. 22) made the first application of the potassium-argon method to an undated mineral. By assuming a value of 3.0 gamma rays per gram of potassium per second to fix the branching ratio, they concluded that lower Oligocene salt beds at Buggingen, Germany, were 20 million years old. Inghram *et al.* (1950, p. 916) determined the amount of radiogenic A^{40} and radiogenic Ca^{40} in a Stassfurt, Germany, sylvite in order to determine a value of the branching ratio of K^{40} which is independent of any assumption regarding the age of the sylvite. Potassium-argon studies by Mousuf (1952) and Russell *et al.* (1953) showed that radiogenic A^{40} yields for microcline feldspars of known ages were consistent with a branching ratio of 0.06. Performing almost the same experiment with different techniques, Wasserburg and Hayden (1954, p. 645) obtained a significantly greater A^{40} yield from one of the samples used by Russell and his coworkers. Repeating the experiment, Shillibeer *et al.* (1954, p. 1794) discovered that by using sodium hydroxide as the fluxing agent instead of sodium as they previously had done their radiogenic A^{40} yields agreed with those of Wasserburg and Hayden who used sodium hydroxide in their check experiment. Wasserburg and Hayden (1955, p. 57) analyzed a series of well-dated potassium feldspars and concluded that the A^{40} yields were consistent with a branching ratio of 0.085. It seemed then that no appreciable argon is lost in feldspars and related minerals and that the constants of the decay process had been determined with enough precision to permit useful geological application. Gentner, Präg, and Smits (1953, p. 11) and Gentner, Goebel, and Präg (1954, p. 124) refined both their techniques and their methods of computation and were able to detect diffusion from certain fine-grained specimens of potassium salts. By postulating an exponential dependence of the diffusion coefficient with time (presumably due to changing temperature) they possibly reconcile argon and helium ages for a variety of these salts. Wetherill, Aldrich, and Davis (1955, p. 171) analyzed a series of feldspars and micas from the same rocks. The rocks varied widely in age and

location. In every case the micas showed significantly greater A^{40}/K^{40} ratios than the feldspars. Folinsbee, Lipson, and Reynolds (1956, p. 60) independently noted the same effect in samples from the Yellowknife area in Canada. The discrepancy cannot be explained simply by diffusion. If simple diffusion from feldspars resulted in lower argon yields as compared to micas, the disparity should become greater as the age of a mineral increases. The percentage of excess radiogenic argon in micas, however, appears to be approximately the same for 200-million-year-old samples as for 2-billion-year-old samples.

Since Aldrich and Nier (1948, p. 876) first extracted radiogenic argon from minerals, improvements enable the experimenter to determine quantitatively the argon content of small, young samples of low potassium content. Aldrich and Nier used samples of the order of 400 grams. To extract the argon they merely heated the mineral to 1000°C. and analyzed the gas with little or no purification. Smits and Gentner (1950, p. 23) released the argon from 200-gm samples by dissolving the salt in water. They then purified the gas by freezing out the water and causing the remainder of the gas to react in a calcium furnace. The amount of argon was measured with a McLeod gauge and collected on chilled activated charcoal for mass spectrometric analysis. Inghram *et al.* (1950, p. 916) first used the isotope dilution technique for the quantitative determination of the radiogenic argon. They incorporated a calcium furnace, a copper-oxide tube, a potassium-hydroxide trap, and a magnesium-perchlorate trap into the purification system. A flux was needed to release the gas from minerals such as feldspars if the process was to be carried out at normal furnace temperatures. Mousuf (1952, p. 150) and Russell *et al.* (1953, p. 1223) used sodium at 850°C. as the flux. Wasserburg and Hayden (1954, p. 645) showed that the sodium flux gave much lower argon yields than a sodium-hydroxide flux at 600°C. Wasserburg and Hayden (1955a, p. 52) outgassed the glass walls of the extraction system by heating with an air torch until a pressure of less than 10^{-6} mm of Hg was achieved.

The principal improvements used at the Berkeley laboratory are the use of the Alpert (1953, p. 860) vacuum techniques in the purification and extraction system, whereby pressures less than 10^{-8} mm of Hg are achieved in the extraction equipment, and the use of a very high sensitivity mass spectrometer especially designed for rare-gas work. It is believed also that the best features of other systems have been incorporated. The refined techniques described in this paper enable the precise dating of a 1-gram sample of Miocene glauconite. Another recent development, not used in the present study, is the use of an induction furnace in the extraction process to fuse minerals such as feldspars without the use of a flux (Carr and Kulp, 1955, p. 380). The relative merits of direct fusion and fusion with flux remain to be established.

In physical measurements of the K^{40} nuclear constant, when K^{40} decays to Ca^{40}, the beta ray emitted in the decay process is comparatively easy to detect. Consequently the decay constant for this mode of decay of K^{40} has not shown wide fluctuations in the experimental determinations in recent years. Beta-ray-decay-rate determinations by 10 laboratories have been averaged (Shillibeer and Russell, 1954, p. 683) to give a mean value of 29.4 beta emissions per second per gram of potassium with an uncertainty of about 6 per cent. Since beta emission is about 10 times more prevalent than decay by K capture, the above constant for the rate of beta decay determines to a large extent the magnitude of the total decay constant. The K capture process which results in A^{40}, on the other hand, does not produce a detectable nuclear-decay particle. An X ray is given off, however, when an electron from one of the outer shells falls into the internal shell vacancy. In turn 88 per cent of these X rays eject Auger electrons (Compton and Allison, 1935, p. 488). These electrons may be counted (Sawyer and Wiedenbeck, 1950, p. 492), but the experiment is difficult because of the low rate of occurrence and the low energy of the process. Since the gamma ray of 1.46 Mev from K^{40} has proved to be in that branch of the decay leading to A^{40} (Johnson, 1952, p. 1213), it should be possible to compute an accurate branching ratio by counting these more easily detectable gamma rays. The average of 11 such determinations made since 1945 gives a branching ratio of 0.110 (Folinsbee *et al.*, 1956, p. 61). When Mousuf (1952, p. 151) first determined geologically a branching ratio of 0.06, the discrepancy with the available counting data was alarming. The solution of the flux problem and the subsequent raising of the geologically determined branching ratio to 0.09 (Shillibeer *et al.*, 1954, p. 1794) did much to reconcile the two types of measurements. Discovery of the mica-feldspar effect seems at last to have

brought the two types of determination into approximate agreement, but an adequate explanation for the systematically lower argon yields from feldspars is still wanting.

In view of the above history it was believed that a useful step could be taken in dating authigenic sedimentary minerals. Even good relative ages in this field would supply a powerful tool to the geologist. Initially, only authigenic materials should be dated because unless the mineral is crystallized at the time of deposition, the determined age would be an average of the ages of the particles composing the sediment. Only in authigenic sedimentary minerals may it be assumed that past crystal structure has been destroyed, so that escape of previously entrapped A^{40} is probable. The problem of formulating a suitable test for the method arises. The only reliable previously reported datings of sedimentary rocks are the single U-Pb date of the Swedish kolm (Nier, Thompson, and Murphey, 1941, p. 112.); the single potassium-argon date of Gentner *et al.* (1954) for lower Oligocene salt deposits of Germany; and four potassium-argon determinations of minerals by Wasserburg and Hayden (1956, p. 134) which range from Paleocene (60 m.y.) to Cambrian (500 m.y.). All these determinations agree generally with the Holmes B time scale (Zeuner, 1950, p. 331) and fall in the right order.

In order to provide a more severe test of the potassium-argon method and to utilize fully the sensitivity of our equipment, we have dated a series of 13 sedimentary minerals (11 glauconites, 1 feldspar, and 1 sylvite) ranging from Miocene (10 m.y.) to Devonian (300 m.y.). Nine of the glauconites are from a well-classified marine sequence in New Zealand which ranges from Miocene to Cretaceous.

Samples were procured, selected, and purified within the Department of Geological Sciences at Berkeley under the direction of Prof. G. Curtis. A detailed explanation of the process of sample preparation is to be found in a companion paper (Curtis and Reynolds (1958).

Acknowledgments

The author wishes to thank Prof. J. H. Reynolds who directed this research. Prof. F. J. Turner conceived the form of the study and has given valuable assistance. Prof. G. Curtis and Prof. R. E. Folinsbee directed the selection and preparation of samples. Samples were collected by Dr. Grange, Dr. C. A. Fleming, Dr. D. R. Gregg and Dr. E. F. Lloyd. The work on which this paper is based was done at the University of California at Berkeley.

Mr. Peter Wygle separated the glauconite from the source rocks, and Mr. Laurence Beal performed the potassium analyses. The glass-blower, Mr. Morley Corbett, and the machine shop under the direction of Mr. W. M. Brower deserve special credit for the fabrication of intricate technical equipment. Many people in the Physics and Geology Departments willingly helped this research.

Financial aid for the preparation of the glauconite samples and for the analyses for potassium was given by The Geological Society of America. Oak Ridge National Laboratories supplied the argon used in the isotopic tracers. The work has been supported by the U. S. Office of Naval Research.

Experimental Methods

Argon Extraction

The extraction system shown schematically in Figure 1 is constructed of pyrex glass mounted compactly on an insulated transite base. Valves separating the divisions of the apparatus are a Berkeley modification of all-metal vacuum valves developed at the Westinghouse Research Laboratories (Alpert, 1953, p. 867). A bake-out oven can be lowered over the major part of the apparatus as indicated by the dotted outline of oven no. 2 (Fig. 1). The vacuum requirements are met by a two-stage mercury diffusion pump mounted beneath the table which supports the entire system. Because the samples require a lower bake-out temperature and the NaOH furnace a higher temperature, these two elements are mounted outside the main bake-out oven. The sample is baked out by a smaller oven (oven no. 1) which may be swung into position when needed. Heating is achieved in oven no. 2 by two 1000-watt strip heaters, controlled by a Partlow temperature regulator. Oven no. 2 is insulated by glass wool packed between two aluminum shells. The sample oven is heated by exposed nichrome wire, and the sodium-hydroxide furnace is heated by exposed nichrome wire mounted in ceramic.

The argon in a sample is extracted and purified in the following manner (Fig. 1): The ground, sieved, and decontaminated sample is split into two parts with a sample splitter. One part is set aside for potassium analysis. The other part is weighed and placed in glass bucket *A*. Fifty grams of sodium hydroxide are placed in the nickel crucible *N*. A vacuum seal is made between the crucible and the system by an aluminum gasket (Fig. 2). Baker purified-calcium turnings are placed in the calcium furnace by filling a small stainless steel crucible which is lowered into a quartz tube on which is wound 25-mil platinum wire.

The isotopic tracer or spike which contains a known amount of A^{38} is sealed onto the system with an iron slug above a break-off seal. These isotope dilution tracers are prepared in groups of 25 by the method described by Balestrini (1954, p. 1503). Each group of spikes has been calibrated by taking

two of the spikes in which the isotopic ratios are known and mixing each of them with an amount of atmospheric argon large enough to be accurately measured in a McLeod gauge. The ratio of the amount of argon in the tracer to the amount of atmospheric argon is then determined by mass-spectrometric analysis.

10^{-8} to 10^{-7} mm of Hg would introduce such a large amount of atmospheric argon that the radiogenic effect would be masked to a high degree.

Valves V_1 and V_2 are closed to isolate the reaction volume. By magnetically rotating iron slugs B the sample is poured into the sodium-hydroxide furnace. Next the A^{38} tracer is released by dropping

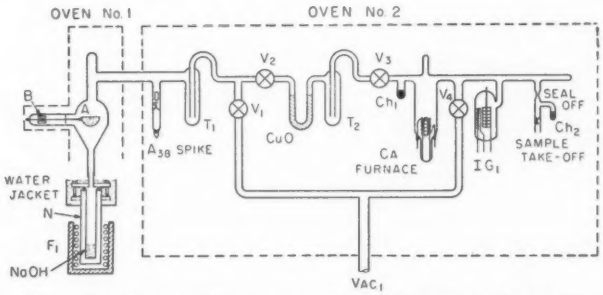

FIGURE 1.—SCHEMATIC DIAGRAM OF ARGON-EXTRACTION APPARATUS

The two mass-spectrometric determinations of the amount of argon in the tracers agree typically within about 3 per cent. This is taken as the probable error in the amount of A^{38} in the tracers.

Also sealed onto the system before each run are: (1) a sample take-off consisting of a constriction, so the purified argon sample may be sealed off under vacuum; (2) a break-off seal, so the sample of gas may be broken open to be introduced to the mass spectrometer; and (3) about 10 grains of activated charcoal 2 mm in diameter to adsorb the gas.

Before the reaction between the sodium hydroxide and the sample is begun, the system is baked out for about 12 hours in the following way: oven no. 1 is maintained at 200°C.; oven no. 2 is maintained at 300°C.; the sodium hydroxide in N is brought to the reaction temperature 600°–650°C. with furnace F_1; finally, the aluminum gasket sealing N is baked out at 100°C. This is accomplished by plugging the run-off tube shown on the left of Figure 2, so the constant-level water control, shown on the right, admits water only at a rate sufficient to replace that lost by boiling. The purpose of the mild bake out of the sample is to remove, in so far as possible, atmospheric gases adsorbed on the surface of the sample. Lepidolite samples, vacuum baked for 72 hours at 200°C. in our laboratory, showed no measurable loss of radiogenic argon, so loss of radiogenic argon during sample bake out, as described above, is thought to be negligible.

Bake out usually lasts overnight, but shorter bake outs may suffice. After the system has cooled, the ovens are raised, and the pressure is measured with the Bayard-Alpert ionization gauge IG_1 (Alpert, 1953, p. 862). If the pressure is 10^{-8} mm of Hg or lower, the run is continued. If the pressure is greater than 10^{-8} mm of Hg, leak hunting is carried out by painting suspect areas with acetone (Alpert, 1953, p. 874). If a leak found in this way can be sealed with glyptal, the run is continued, but the leak must be permanently repaired before the next run. It should be explained that the principal reason for the bake out is to check for leaks in the system. A leak which would degrade the pressure from

an iron slug on the break-off seal. Now the reaction between the sample and the sodium hydroxide is started by slowly raising the temperature of the nickel crucible N to 600°–650°C. A slow increase in temperature is especially important in the case of glauconites, because this mineral tends to react very violently with the sodium hydroxide. To facilitate this process a variac drive was designed which turns a standard variac from 2.5 to 10 dial units an hour, depending on the setting of a variable resistance in series with the field coils of the dc motor which powers the drive. During the reaction the water bath is changed to a cold-water bath by opening the run-off tube which was previously plugged. Keeping the water bath cold was found to prevent creep of the sodium hydroxide in the nickel crucible.

The crucible is kept at maximum temperature for 12 hours to allow the tracer to reach equilibrium with the released gas from the rock sample. Often large quantities of water are released by the reaction. If this happens trap T_1 is chilled with liquid nitrogen to collect this water and prevent it from settling in the valves. While the sample is reacting the copper oxide furnace, *CuO*, is outgassed at its subsequent operating temperature of 500°C., and the charcoal traps Ch_1 and Ch_2 are given a final bake out at 450°C. The charcoal traps are kept warm until used to prevent the adsorption of unwanted gas. Calcium vapor is distilled from the crucible onto the pyrex walls of the calcium furnace. The mirror thus formed is moved about by manipulation of an air torch. Finally any gas that may have been adsorbed by the gauge during the other outgassing processes is eliminated by heating the gauge electrodes.

The gas evolved is first purified by passing the gas over the copper oxide at 500°C. Hydrogen and carbon monoxide are thereby converted to water and carbon dioxide which are subsequently frozen out by trap T_2. In many cases T_1 is not used, and the original water of reaction is also trapped in T_2. This purification is allowed to continue for at least half an hour at maximum copper-oxide temperature.

The remaining gas to be purified is transferred to

the calcium furnace by chilling Ch_1 with liquid nitrogen for about 20 minutes. The gas is released from the charcoal, and the calcium purification is initiated. Calcium vapor is produced, and the subsequent deposit is moved about until a considerable amount of unreacted calcium is visible on the walls

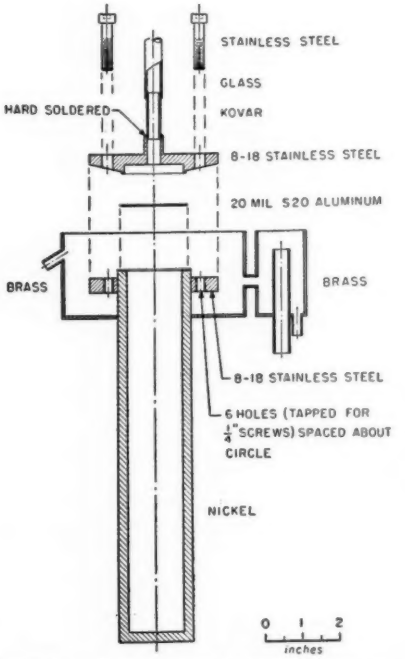

FIGURE 2.—SCALE DRAWING OF SODIUM-HYDROXIDE FURNACE

of the furnace. The calcium furnace is then turned off and allowed to cool. Liquid nitrogen is employed to chill Ch_2; the purified argon is absorbed on the charcoal, and the trap is kept chilled for at least 10 minutes. Finally the sample is sealed off at the constriction with an oxygen torch.

Mass Spectrometric Analysis of Argon Sample

Figure 3 is a schematic diagram of the sample-analysis system. The mass spectrometer is a high-sensitivity instrument especially designed for rare-gas analysis, which has an all-glass envelope and gives a low background (Reynolds, 1954, p. 283; Lipson and Reynolds, 1954, p. 283). The spectrometer has a 4½-inch-radius 60°-magnetic-sector analyzer and an electron-bombardment ionization source. The signal is detected by a vibrating reed electrometer with a 5×10^{10}-ohm input resistor. The output of the electrometer is recorded on a Brown strip-chart potentiometer. By baking the mass spectrometer overnight at 375°C., the pressure is reduced from atmospheric to less than 10^{-9} mm of Hg. Under operating conditions no peaks are visible at masses 36, 38, and 40. Background is kept at a minimum by continuous pumping between runs, and further bake out is not necessary until after the apparatus has been raised to atmospheric pressure for maintenance purposes. The background at mass 36 is the most critical since the height of the small A^{36} peak determines the atmospheric argon correction.

After the sample is sealed onto the sample-introduction system, this volume to the left of valve V_8 is evacuated but not baked out. When a pressure of 5×10^{-6} mm of Hg is reached, the sample can be analyzed. During all mass-spectrometer operations the ionization gauge IG_3 is prevented from performing any pumping by floating the filament (Alpert, 1953, p. 875). The background-mass spectrum is first checked with valve V_7 open. This valve is then closed thereby isolating the mass spectrometer. The background can now build up since there is no pumping action on the volume of the mass spectrometer. If the run is being conducted shortly after another argon analysis, a memory effect will be seen in which the argon isotopes of mass 38 and 40 build up roughly in the ratio of the last samples to be analyzed. This effect can be diminished by further pumping on the mass spectrometer or eliminated by baking out the mass spectrometer. Once the background has been deemed satisfactory, the actual analysis of the sample is conducted. Valve V_5 is closed, and the break-off seal of the sample is broken. The sample thus escapes into the volume between valves V_5 and V_4. The mass spectrometer may be operated by a static method or by the more conventional flow method. In the static method of operation, a small amount of sample is introduced to the mass spectrometer by slightly opening valve V_8 for a short time. Valve V_8 is then closed again, and a small amount of sample is trapped in the volume of the mass spectrometer. In order to allow the gas to enter the body of the mass spectrometer freely, a small metal ball valve, *MBV*, is then moved away from a hole with a ground glass seat. This hole is shown just to the left of the metal ball in Fig. 3.

The advantage of the static method is that extremely small amounts of sample can be carefully and leisurely analyzed. Pumping action by the mass-spectrometer beam or by the surfaces of the instrument is found to be small, so the peak heights decrease very slowly with time. A disadvantage is that the memory effects can be accentuated by the large ion beam of the sample bombarding the walls of the spectrometer tube. Even though the amount of the sample is small, the beam produced is over 100 times the size of the background beam. A second disadvantage is that fractionation occurs when the sample is introduced. The lighter isotopes enter the spectrometer in greater number in proportion to the square root of their masses. As long as the fraction of sample introduced is small, this effect can be corrected for by multiplying the ratios of the peak heights by the square roots of the masses involved.

In the flow method of operation, valve V_8 is regulated to allow a constant stream of gas to enter the mass spectrometer while valve V_7 is left open to pump the gas out. The metal ball valve is moved to cover the hole in the sample entry tube and thus force the gas to flow directly into the ionization box of the ion source. The advantage and disadvantages are the reverse of those of the static method.

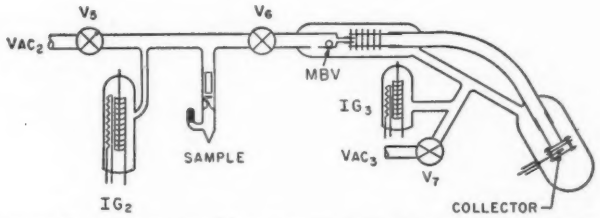

FIGURE 3.—SCHEMATIC DIAGRAM OF THE MASS SPECTROMETER AND THE SAMPLE INTRODUCTION SYSTEM

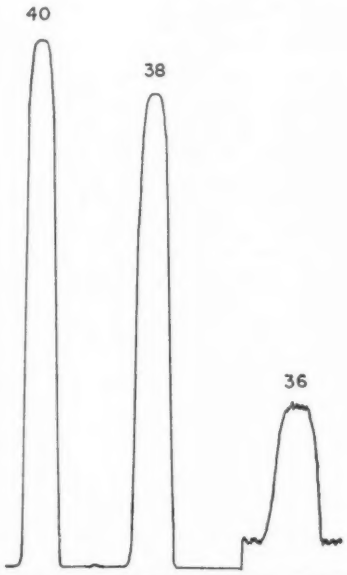

FIGURE 4.—BROWN POTENTIOMETER TRACE OF MASS SPECTROMETER OUTPUT IN A SWEEP OVER MASSES 36, 38, AND 40

The sensitivity of the electrometer in recording the A^{36} peak is 100 times that used in recording the A^{38} and A^{40} peaks.

The mass spectrum is analyzed by sweeping the magnetic field while the accelerating voltage is held constant. Figure 4 shows the Brown potentiometer trace of the vibrating reed-electrometer output in a typical sweep over the masses 36, 38, and 40. After the 40/38 ratio has been determined by at least 5 sweeps over these masses, more sample is introduced to magnify the A^{38} and A^{36} peaks. The ratio 38/36 is then determined by at least 5 more sweeps over these masses. When the static run is finished, the spectrometer is pumped out, and the flow analysis is conducted as a check.

Analysis of Data

Since the detected ion beam varies slowly but systematically with time, drift lines are drawn between the peaks of mass 38 and the peaks of mass 40. The peak height measurements for masses 36, 38, and 40 are made at the position of mass 36 on the chart, and the heights assigned to masses 38 and 40 are the distances from the extended base lines to the drift lines.

After the raw peak heights are tabulated ratios 40/38 and 38/36 are computed. These ratios are then averaged and, the deviations and average absolute deviation are determined. The corrected (square root of mass correction) static data are compared to the flow data and are found to agree. Memory effects in the static method can be detected by a steady change of the ratios with time if the effects are appreciable. This can be corrected by the extrapolation of the ratios to zero time, *i.e.*, the time when the sample was first admitted to the spectrometer. Finally a small (about 1 per cent) discrimination correction is applied. In determining the discrimination correction, abundance ratios of atmospheric argon as determined on our instrument are compared with those of Nier (1950, p. 792). The discrimination arises chiefly because of the presence of stray magnetic field from the magnet at the ion source. This is one disadvantage of mass spectrometers of small radius.

Determination of Potassium

The potassium content of the samples was determined by use of a Perkin-Elmer flame photometer using a lithium internal standard. Potassium determinations on National Bureau of Standards standard potassium feldspar (No. 70) were interpolated in the sequence of samples. Our results always were within 1 per cent of the gravimetric data for this feldspar. A further check has been supplied by two samples which were analyzed elsewhere by completely different methods: an olivine basalt at the 0.x per cent potassium level where potassium was determined gravimetrically and a meteorite at the 0.0x per cent level where potassium was determined by isotope dilution. In view of the above checks and the consistency of repeat runs, it is believed that the flame-photometric potassium analyses reported here are accurate to within a probable error of 3 per cent.

CALCULATIONS

The fundamental equation used in calculating the age of a potassium-bearing mineral is

$$t = \frac{1}{\lambda} \ln \left[1 + \frac{A_r^{40}}{K^{40}} \left(\frac{1+R}{R} \right) \right]$$

where

t = age in years,
λ = total decay constant of K^{40} per year = $\lambda_\beta + \lambda_K$
A_r^{40} = number of moles of radiogenic argon in the sample
K^{40} = number of moles of K^{40} in the sample
R = branching ratio = λ_K/λ_β

The decay constant λ_β, giving the number of decays per year per atom of K^{40} to calcium, is determined to be $\lambda_\beta = 0.503 \times 10^{-9}$/year (Shillibeer and Russell, 1954, p. 685). Therefore as soon as a satisfactory branching ratio is decided upon, we can assign the total decay constant. As has been mentioned above, experiments with pegmatitic samples, for which reliable uraninite ages are available, have indicated that use of a branching ratio of about 0.09 is highly successful in bringing the ages of potassium-rich feldspars into agreement with uranium-lead ages. Micas, however, give correct ages when a higher branching ratio, approximately equal to the physically determined branching ratio, is employed. Since the samples under study are not of known age but only of known order in age, it is impossible to assign an empirical branching ratio. For this reason we have employed the physically determined branching ratio computed by averaging values determined within the last 12 years for the specific gamma activity of potassium and computing the ratio of this figure to the well-known specific beta activity. An average of 11 determinations yields a value of 0.110. The total decay constant λ is then found to be

$$\lambda = 0.558 \times 10^{-9}/\text{year}$$

The number of moles of radiogenic A^{40} is determined from the isotopic analysis of the argon. If we let

R_{38}^{40} = the ratio of A^{40} to A^{38} in the gas sample
R_{36}^{38} = the ratio of A^{38} to A^{36} in the gas sample
N_s = the number of moles of argon in the tracer
A_r^{40} = the number of moles of radiogenic A^{40} in the mineral
f_s^{38} = the atom fraction of A^{38} in the tracer
s_{38}^{36} = the ratio of A^{36} to A^{38} in the tracer
s_{38}^{40} = the ratio of A^{40} to A^{38} in the tracer
a_{38}^{36} = the ratio of A^{36} to A^{38} in air
a_{40}^{36} = the ratio of A^{36} to A^{40} in air
a_{40}^{38} = the ratio of A^{38} to A^{40} in air

then

$$A_r^{40} = f_s^{38} N_s \left\{ R_{38}^{40} \left[1 + \frac{1 - s_{38}^{36} R_{36}^{38}}{R_{36}^{38} a_{36}^{36} - 1} \right] - \frac{1 - s_{38}^{36} R_{36}^{38}}{R_{36}^{38} a_{40}^{36} - a_{40}^{38}} - s_{38}^{40} \right\}$$

The above equation applies to any mixture of radiogenic, atmospheric, and tracer argon. If no sample has been introduced to the extraction apparatus the calculated amount of A_r^{40} should be zero. This was applied to a blank run of the extraction system in which all steps were carried out as if a sample were present, but no sample was introduced. The result was that the apparent radiogenic argon was much less than 1 per cent of the amount of the tracer and thus constituted an over-all check of the computation procedure.

If we let

K^{40} = the number of moles of K^{40} in the sample,
f_k = the fraction by weight of potassium in the sample,
S = the number of grams of sample,

then

$$K^{40} = Sf_k(0.000119)/39.100$$

where

0.000119 = the atom fraction of K^{40} in potassium (Nier, 1950, p. 793)
39.100 = the atomic weight of potassium

The A_r^{40}/K^{40} ratio may now be computed with a probable error of from 5 per cent to 7 per cent, depending upon the percentage of atmospheric argon in the sample. The total probable error arises from a probable error of 3 per cent in the potassium determination and a probable error of 4 per cent in the radiogenic argon determination exclusive of error in the atmospheric argon correction. The error in the determination of the A^{36}/A^{40} abundance ratio is taken to be 3 per cent, but the effect of this error on the radiogenic argon determination varies with the percentage of atmospheric argon. If we let

e = the percentage error in A^{36}/A^{40},
f = percentage of atmospheric argon in the sample,
E = the percentage error in the radiogenic argon due to the error in A^{36}/A^{40},

then

$$E = \frac{ef}{100 - f}.$$

Applying this equation by using the percentages of atmospheric argon which are tabulated for each sample, the total probable error can be calculated. In addition, sample 39 is assigned a greater probable error (7 per cent) than is called for at first glance, because trouble with the electronic equipment during the static run forced the beam to be operated for a long time before satisfactory sweeps were obtained.

Possible errors are: (1) incomplete recovery of argon; (2) loss of tracer argon before it is in equilibrium with the argon released from the sample; and (3) an impurity contribution to the mass 36 peak which is not visible in the background before the sample is introduced. The recovery of argon by fused sodium hydroxide has been checked against variation with grain size (Shillibeer and Russell, 1954, p. 689). At this laboratory a sample was rerun at 800°C. (sodium-hydroxide temperature) to check against variation of argon yield with temperature (Folinsbee *et al.*, 1956). There was no additional argon recovery.

In order to lose 1 per cent of the tracer in 12 hours, the valves sealing off the reaction must have a leak rate greater than 10^{-7} liters/sec. The all-metal valves normally have a leak rate of less than 10^{-9} liters/sec. Those used in this experiment have been checked periodically and have shown no deterioration of the leak rate.

The possibility of an impurity contribution to the 36 abundance can be ruled improbable by examination of the mass spectrometric record. The main concern at mass 36 is hydrochloric acid and hydrocarbon contamination. Therefore the mass positions other than 36, 38, and 40 are observed for the appearance of unexpected peaks and for changes of peak height upon introduction of a sample. Small Cl^+ background peaks at masses 35 and 37 are the result of the process used to make the spectrometer tube conducting, but there are no background peaks at masses 34, 39, 41, 42, or 43. A change in height at masses 35 and 37 or the appearance of a peak at masses 34, 39, 41, 42, or 43 would suggest that impurities were being introduced which might result in a contribution to mass 36. These changes did not occur in the runs reported here. A further check occurs when the 38/36 ratio is computed for different amounts of sample in the spectrometer and for the different modes of spectrometer operation. If the 38/36 ratio remains constant under varying conditions, the result is considered more reliable.

Results

Figure 5 is a map of New Zealand showing the location of the samples. Results are tabulated in Table 1. The column giving the specific gravity is only pertinent to glauconites. The blank determination shows that most of the argon comes from the mineral samples and not from the apparatus.

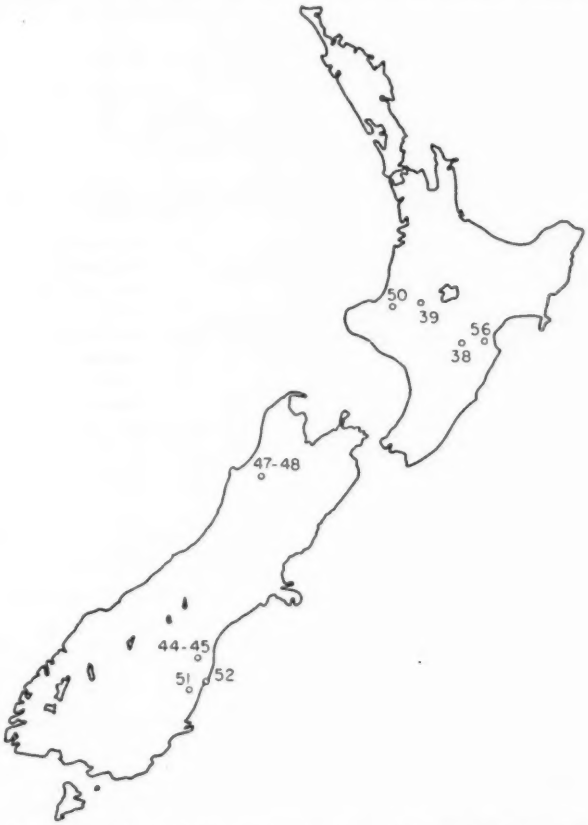

FIGURE 5.—MAP OF NEW ZEALAND SHOWING LOCATION OF SAMPLES

Figure 6 is a graph of assigned age versus determined age for the New Zealand samples. Ages are assigned by interpolation on the Holmes B time scale (Zeuner, 1950, p. 331). Included for comparison are the age of the lower Oligocene sylvite as determined by Gentner *et al.* (1954, p. 124), the age of a Paleocene (possibly upper Cretaceous) glauconite from Canada as determined by Wasserburg and Hayden (1956, p. 134) (recalculated using $R = 0.110$ and $\lambda = 0.558 \times 10^{-9}$/year). Also included is the Colorado, U. S. A., pitchblende age which was determined by Nier *et al.* (1941, p. 113) and establishes the date of the end of the Paleocene.

Samples 44 and 45 were taken about a quarter of a mile apart from the same faunal stage. Sample

38 is from the boundary between the Paleocene and Cretaceous. Samples 47 and 48 are different sieve fractions of the same original sample. These (47 and 48) are the most contaminated samples that were run, and it is believed that the source material of the contamination could be as old as 500 million years. All the samples analyzed had less than 1 per cent contamination by volume except the two samples from the Tongaporutuan, 47 and 48. Sample 47 had 4 per cent contamination by volume, and sample 48 had 1 per cent contamination by volume. All the New Zealand samples are surface samples, except sample 56 which is a core sample.

TABLE 1.—EXPERIMENTAL VALUES

Sample	Weight gms	Sieve fraction	Specific gravity	%K	% atm. argon	atm. A per gm sample, moles x 10^{10}	(A^{r}_{40}/K_{40}) x 10^{3}	Stage	Age years x 10^{-6}
New Zealand Glauconites									
47	4.33	48-65	< 2.65	3.42	54	2.5	2.00	Tongaporutuan	35.8±2.1
48	3.50	100-150	< 2.56	3.32	68	3.8	1.71	Tongaporutuan (same rock as 47)	30.2±1.8
56	1.107	48-65	< 2.65	4.80	64	3.5	1.17	Waiauan	21.1±1.5
52	3.23	48-65	< 2.55	5.56	48	1.5	0.905	Hutchinsonian	16.3±1.0
50	0.722	48-65	> 2.65	5.12	55	2.7	1.14	Waitakian	20.4±1.2
39	5.80	35-65	< 2.65	5.93	62	3.7	1.22	Duntroonian-Whaingaroan	22.0±1.5
44	2.07	16-48	> 2.66	6.11	69	9.0	2.07	Kaiatan	37.0±2.6
45	0.875	16-48	> 2.65	5.87	37	2.6	2.26	Kaiatan	40.5±2.8
51	5.72	48-65	< 2.65	5.82	48	4.6	2.76	Waipawan	49.2±3.0
38	6.28	65-100	< 2.65	6.82	58	5.7	2.58	Just below Waipawan	46.1±2.8
Feldspar from Crowsnest Volcanics, Alberta, Canada								Epoch	
22	4.32	20-65		10.82	6	1.1	5.40	Cenomanian	94.9±4.7
Glauconite from MacMurray Area, Canada (Socony Vacuum Oilsand Well No. 27, Twp 19 Rge 10 Scd 27 W5)									
41	2.54	65-200	< 2.65	4.01	26	3.6	8.24	Albian	142±10
Sylvite from top of Elk Point Formation, Saskatchewan, Canada									
32	3.13	16-20		4.20	6	1.3	17.1	Givetian	285±14
Blank						0.5 for assumed 1 gm sample			

Figure 7 incorporates three older samples from Canada (22, 41, 32), an older Canadian glauconite of Wasserburg and Hayden (1956, p. 134) and a samarskite age of Nier *et al.* (1941, p. 113) which was used by Holmes to establish the Devonian-Carboniferous boundary. The age of the glauconite of Wasserburg and Hayden (1956, p. 134) was recalculated using the same decay constants as the other samples reported here.

DISCUSSIONS AND CONCLUSIONS

Except for the three determinations from the Miocene, the New Zealand sequence of glauconites presents a remarkably linear plot when the uncertainties in the assigned ages are considered. The two samples from the same stage, 44 and 45, agree within the assigned probable error. Nevertheless the departure of the Miocene samples is disturbing. The high age for the Tongaporutuan sample is probably due to contamination, but the Waiauan sample is exceptionally free from potassium-bearing contaminant. Although the gross contamination was fairly high in the Waiauan sample, about 90 per cent of the contaminating material was quartz.

Opposed to this discrepancy is the agreement with two recent determinations by the potassium-argon method. Gentner *et al.* (1954, p. 124) determined the age of lower Oligocene salt deposits at Buggingen, Germany, to be 25_{-3}^{+4} million years, using almost the same decay constants as those used in this paper. This figure was derived from the original age determination of the same sylvites which was 20 million years. It is computed by showing that the variation of argon content with crystal size is accounted for by diffusion and by bringing argon and helium ages into agreement by assuming an exponential decay of the diffusion rate by a cooling down of the salt.

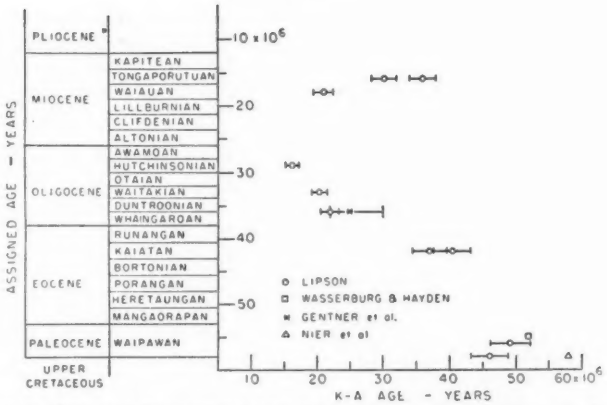

FIGURE 6.—PLOT OF TERTIARY POTASSIUM-ARGON AGES VERSUS AGES ASSIGNED BY INTERPOLATION IN THE HOLMES B TIME SCALE

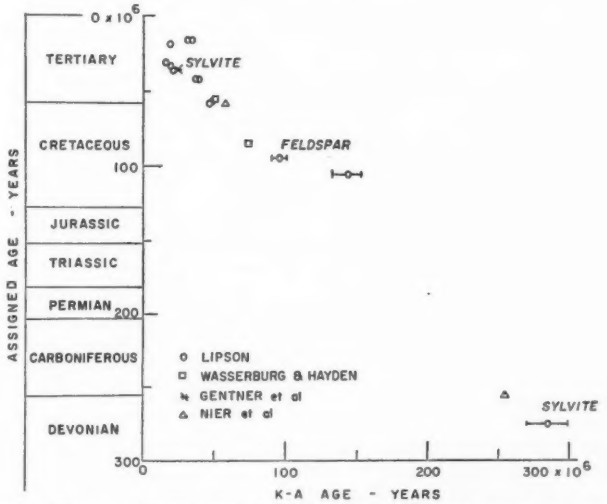

FIGURE 7.—PLOT OF POTASSIUM-ARGON AGES VERSUS AGES ASSIGNED BY INTERPOLATION IN THE HOLMES B TIME SCALE

Gentner *et al.* (1954, p. 124) feel that within the limits stated the age is definitive. Curtis and Reynolds (1958) question the validity of the calculation, however.

At the lower end of the range of the New Zealand samples, agreement is found with a Canadian glauconite sample recently analyzed by Wasserburg and Hayden (1956, p. 134) who believe that the sample is most probably Paleocene but possibly Upper Cretaceous. The age of the Upper Paleocene pitchblende was determined by Nier *et al.* (1941, p. 113) from the ratio U^{238}/Pb^{206}. The common lead correction was large and uncertain. The only completely trustworthy lead ages are those for which the U^{235}/Pb^{207}, U^{238}/Pb^{206}, and Pb^{208}/Pb^{207} ages agree, providing an internal check that the U-Pb system in question was really a closed system during the whole lifetime of the sample. In view of the general lack of trust in a uranium-lead age for which such an internal check is lacking, and the lack of knowledge about argon loss from glauconites, the discrepancy from the Nier age for the Paleocene is not alarming. The internal agreement with two

potassium-argon dates by other workers, however, is encouraging.

Continuing into the Cretaceous, the samples fall in the right order, but sample 41 is certainly too old. The MacMurray glauconite (Albian) should be only slightly older than the feldspar from the Crowsnest volcanics (Cenomanian—95 m.y.) The sylvite (sample 32) falls precisely in its proper place on the Holmes scale. The Nier *et al.* (1941, p. 113) samarskite age for the Devonian-Carboniferous boundary which appears in Figure 6 is from a sample in which the U-Pb ages agree.

Since the results are on the whole reasonable, it seems worthwhile to attempt to explain the three samples which appear too old when compared with the other potassium dates at our disposal. The high age for the Miocene samples from the Tongaporutuan and the Albian sample may both be the consequence of contamination from older material embedded in the glauconite. To test this a comparatively pure sample of nonglauconitic mineral from these and from other samples of the same areas should be separated and dated.

The relatively high age for sample 56 must be explained on some other basis. Sample 56 is the only New Zealand sample which is a core sample. All the other samples are surface samples and are slightly weathered. The core sample, not being exposed to the weathering process, has possibly retained relatively more argon. A sequence of uncontaminated core samples should be investigated to see if they are uniformly older than a sequence of uncontaminated surface samples. The Albian glauconite is also a core sample.

These three samples have the lowest potassium content of all the samples dated. Low potassium content possibly has some relation to crystal structure, and crystal structure in turn may determine how well the mineral can retain argon. X-ray patterns of low- and high-potassium glauconites in combination with the dating of more low-potassium samples should resolve this point.

The unmistakable amounts of atmospheric argon found in the glauconite samples (Table 1, column 7) suggests that they crystallized in an environment not free from argon. This argon may have been a mixture of atmospheric and radiogenic argon. Certainly our results do not eliminate this as an explanation, but the regularities found are evidence against it.

The results are encouraging enough to warrant continued potassium-argon dating of sedimentary rocks. It is too early to propose an absolute time scale for the Tertiary, and whether the apparent discrepancy between results for the Miocene and Oligocene samples arises from an overestimate of the age of the former or an underestimate of the age of the latter is not known. As an increasing number of carefully chosen samples is dated, either a valuable tool for determining relative stratigraphic age will be found or, failing in this, some insight into the geological history of sedimentary deposits will be gained.

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NOTES ON THE POTASSIUM-ARGON DATING OF SEDIMENTARY ROCKS

BY G. H. CURTIS AND J. H. REYNOLDS

ABSTRACT

Lipson's companion paper on the potassium–argon dating of sedimentary rocks is discussed. Some limitations in the present geological time scale are considered. The sedimentary minerals to which K-A dating may be applied and methods used in the preparation of glauconite for analysis are described. Possible errors due to contamination, argon inheritance, and argon loss by diffusion are discussed. Evidence by Gentner and co-workers for argon diffusion in sylvite is reviewed critically.

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INTRODUCTION AND ACKNOWLEDGMENTS

Geologists have been keenly interested in physical methods of age determination which might be applied successfully to sedimentary rocks. The few calibration points of the geologic column depend for the most part on physically determined dates of igneous rocks which are difficult to place precisely in the column.

From the time of its discovery (Aldrich and Nier, 1948), the potassium-argon method has been viewed as a possible key to the dating of sedimentary rocks since the parent element, potassium, is an important constituent of several sedimentary minerals. The dating of igneous minerals has been tested many times within the past 5 years with success. It has now been shown, for example, that igneous micas ranging in age from 1 million (Curtis, Lipson, and Evernden, 1956, p. 1360) to thousands of millions of years (Wetherill, Aldrich, and Tilton, 1956, p. 362) contain detectable radiogenic argon in the amounts to be expected from the potassium contents, the ages of the micas as established by other methods, and from the physically determined decay constants for K^{40}. Improvements in techniques, such as those described by Lipson (1958), make it possible to work with small samples, *i.e.* of about 1 gm, even when the material is relatively recent and contains only a small percentage of potassium. These facts indicate that K-A dating can be applied on an exploratory basis to many geological problems involving sedimentary rocks. The companion work by Lipson (1958) represents such an application.

The authors take pleasure in acknowledging helpful discussions with Professors John Verhoogen and Charles Kittel. The potassium-argon dating program has received financial aid from The Geological Society of America. This work has been supported by the U. S. Atomic Energy Commission.

PROBLEMS IN TIME CALIBRATION OF THE GEOLOGIC COLUMN

One of the goals of geochronometry is to establish precise dates in years for as many

stratigraphically identified events as possible; the ultimate aim is to establish the sedimentary record as an accurate time chart. Currently the best progress in this direction is in the time scales constructed by Holmes (1947, p. 145). In these scales four of the five calibration points are uranium-lead dates for intrusive igneous rocks or veins. Unfortunately the physically determined age for an intrusive igneous rock or associated veins rarely can be correlated precisely to the geological column. The basis for such a correlation is provided when the igneous rock intrudes sedimentary beds of known geologic age and is unconformably overlain by younger beds of known geologic age. Certainly the physical date for the rock will lie between the true dates of deposition of the two layers since the igneous rock will have cooled after the first event but before the second. Orogeny, granitic intrusion, uplift, erosion, and submergence, however, are not accomplished in a geological moment. Knopf (1949, p. 4) and Moore (1952, p. 151) noted that in four of the cases used by Holmes the events cannot be dated geologically more closely than a geologic epoch; in some cases, even the period is in question. Besides the uncertainty of the geologic age in such instances is the question of whether the physical age concurs with the true time of emplacement of the granitic body being dated. Plutonic igneous rocks, whether formed by granitization or crystallization of a molten magma, probably remain warm after consolidation for protracted periods of time, perhaps millions of years. During crystal growth in a plutonic environment, which probably spans more than 1 million years (Lovering, 1955, p. 277), radiogenic elements not needed in the crystal lattice will probably diffuse out as rapidly as formed. During the period of slowly declining temperatures, after completion of the major part of crystal growth, diffusion of radiogenic elements out of the lattice may keep pace with their formation for a very long time. The length of the period of diffusion must remain a matter for speculation until more is learned about plutonic processes and the depth and temperatures at which they occur. Larsen (1945, p. 408) concluded that it would take approximately 7 million years for a large batholith to cool from 800°C. to 200°C. at a depth of 5 km. Lovering (1955, p. 278) points out that this time might be greatly extended if the batholith were composed of separate plutons successively emplaced, a characteristic of most batholiths. Throughout much of the 600° temperature drop diffusion of the radiogenic elements could be rapid. Thus the birth date determined for a plutonic mineral would be near the time when the diffusion rate no longer exceeded the radiogenic-accumulation rate and could be several million years after the true birth date, the time of crystallization of the mineral.

It is especially dangerous with our present knowledge to attempt to use a geologic time scale, such as the Holmes scale, together with a physical date for an igneous rock to identify the precise geologic epoch or stage in which the corresponding pluton was emplaced because both the calibration points on the time scale and the true age of the igneous specimen are uncertain. For example, dates of from 100 to 105 million years for the Sierra Nevada, the Idaho, and the Southern California batholiths (Larsen *et al.*, 1954, p. 1277) fall approximately in the middle of the Cretaceous on the Holmes time scale B, but the uncertainties involved make it dangerous to assign the emplacements to any interval less comprehensive than one extending from the Portlandian of the Upper Jurassic to the Cenomanian stage of the Upper Cretaceous.

The differences between true and apparent ages of minerals is of little consequence in considering ancient plutonic masses, as they both will probably fall within the limits of error of all present methods of age determination.

The geologic significance of physically determined ages of regionally developed metamorphic rocks (where temperatures are known to be high at the outset) and contact metamorphosed rocks adjacent to large plutonic bodies may be held in question for similar reasons. There are two geologic environments in late geologic history, however, (Cretaceous to Recent) in which minerals whose determined ages may be reasonably close to their true ages have crystallized, and hence may be suitable for placing geologic events within a particular epoch or stage with some degree of accuracy. These are the extrusive igneous (or shallow intrusive) and sedimentary environments.

Extrusive igneous rocks—domes, flows, and tuffs—cool within a matter of years at most; if they contain suitable minerals which have not suffered alteration subsequent to their formation they should be satisfactory for dating by physical methods. It seems safe to predict that the ultimate geological time chart will be based for the most part on physical dates for such minerals. The sedimentary environment, although probably more complex than any of the other environments of mineral formation, favors certain short-term processes of mineral

growth at the water-sediment interface. Suitable minerals formed here may thus be used to date the period of sedimentation.

Sedimentary Minerals Suitable for K-A Dating

The detrital grains composing most sedimentary rocks are, as a rule, unsuitable for age determinations if the purpose is to date the period of sedimentation. Detrital grains crystallized in other geologic environments in previous geologic periods, and their ages have no significance in the formation of the rock in which they now occur. The mineral grains composing volcanic crystal tuffs are exceptions to this rule. Where suitable unaltered volcanic grains, particularly mica and potassium feldspar, can be recognized in sedimentary rocks, they are ideal for dating purposes, as they were probably formed shortly before their introduction to the sedimentary environment.

Authigenic minerals, or those which have crystallized entirely within the sedimentary environment or later in the sedimentary rock itself, must, to be useful for K-A dating, contain potassium, must have grown essentially at the water-sediment interface, and must have retained most of their radiogenic argon. The ubiquitous potassium-rich clays may ultimately satisfy the above requirements; however, it is known that many clay minerals undergo long-term diagenetic changes after deposition, and have probably lost much of their radiogenic argon. Until more is known about their behavior they must be considered poor minerals for dating. There are several salts of potassium found in sedimentary rocks such as aphthitalite, carnallite, douglasite, hanksite, kainite, kaliborite, langbeinite, leonite, picromerite, polyhalite, syngenite, and sylvite. Most of these are rare; their formation is incompletely understood; and, except for sylvite, their ability to retain radiogenic argon is unknown. Sylvite, the most common and best known of these salts, has been used by several workers (Gentner, Präg, and Smits 1953, p. 11; Folinsbee, Lipson, and Reynolds, 1956, p. 65) but its scarcity in most sedimentary sections prevents its wide-scale use. Orthoclase is known to occur as an authigenic mineral in sedimentary rocks, but its period of formation in relation to sedimentation is questionable, and the problem of separating it from associated detrital orthoclase is almost insurmountable.

Glauconite, a complex hydrous potassium-bearing silicate of variable composition, is common in many sedimentary rocks. It seems to meet most of the necessary requirements and in addition it has other desirable attributes. Although their mode of origin is not completely understood, most glauconites probably formed by diagenetic processes at the water-sediment interface and not after a long period of burial. The age is thus probably close to that of the period of accumulation of its associated detrital sediments. The large grain size in comparison to many authigenic minerals and the distinctive green color and ovoid shape makes glauconite easy to find. Owing to its high iron content it is particularly adapted to magnetic separation. Coupled with its large potassium content, which ensures a relatively high amount of radiogenic argon, is its micalike structure, favorable for argon retention.

Sample Preparation

Separating glauconite from detrital sedimentary rocks is generally a simple matter. The rock is crushed, wet-screened (+150, −28 Tyler mesh), dried, and passed through a Franz magnetic separator. Contaminating minerals are removed from the glauconite concentrate with bromoform of appropriate specific gravity.[1] Magnetite should be extracted with a hand magnet before the magnetic separator is used, and calcium carbonate which adheres to the glauconite can be removed with acetic acid. Thin sections are made of every sample before crushing to indicate the amount of contamination within the glauconite grains. Grain mounts are made of the various size fractions after separation to check purity. In order to observe impurities within the grains, grain mounts must be ground to a thickness of approximately 30 microns. The amount of contamination by exterior grains (feldspar, mica) is determined by simple count, and the amount of contamination within the grains is measured with a point counter.

Few data on the role the variable physical and chemical properties of glauconite play in retention of radiogenic argon are available. Glauconite may vary in specific gravity from less than 2.3 to more than 2.75 in a single sample (Smulikowski, 1954, p. 22). It seemed reasonable that higher specific gravity would indicate more compact structure, hence greater

[1] Hessland, Lukins, and Freden (1949, p. 571–578) have developed a dielectric procedure modified after Berg (1936, p. 23–27) for the separation of glauconite. Their technique is probably as effective as that discussed above but seems more cumbersome.

retention of argon. In Lipson's work, however, runs of two different specific gravity fractions from the same sample (−2.45 and 2.65+) did not differ significantly in their ages. Reasoning that diffusion effects would be less in large grains, Lipson made most age determinations on the coarsest glauconite fraction obtainable from a particular sample. Again, however, runs of two different size fractions from the same sample (−100, +150 and +60 Tyler mesh) failed to yield a significant difference.

A minimum of 1–2 gm of pure glauconite is necessary for an argon determination of rocks of mid-Tertiary age or older. Several times this amount should be obtained to ensure enough material for duplicate determinations of both argon and potassium. For rocks as young as Pleistocene approximately 20 gm of material are required. Although glauconite may compose more than 50 per cent of a particular sample , at least 50 gm of sample may be necessary to obtain 3–4 gm of glauconite contaminated with less than 1 per cent of clastic grains, such as feldspar. The deleterious effect of these minerals depends upon their potassium content, their absolute age, and their abundance within the grains of glauconite.

Sources of Error

Types of Error

For age determination, a sample must have certain properties, namely a reasonably definite time of crystal formation followed by a period during which there has been no extensive alteration. It is inevitable that many specimens with which dating is attempted will fail to meet these specifications. The suitability of a specimen for dating purposes is, after all, an opinion based on the geologist's conception of the history of the specimen and is subject to error. Age agreement on several samples by a single method, on a single sample by several methods, or—best of all—both, is the only available means of proof that alteration or recrystallization errors are absent. Since it is the purpose here to discuss sources of error peculiar to the K-A method as applied to sediments, the sample in question is assumed to be correctly evaluated by the geologist as suitable for dating. The distinctive sources of error are contamination, argon inheritance, and diffusion.

Contamination

The problem of contamination is not as serious with mineral concentrates from igneous rocks as it is with mineral concentrates such as glauconite from sedimentary rocks. All the minerals composing an igneous rock are probably approximately the same age. The detrital grains composing a particular sedimentary rock, however, may be of almost any age. When working with young glauconites, of Cretaceous age or younger, contaminating clastic grains of orthoclase derived from ancient terranes are potentially a serious source of error. Orthoclase contains approximately 2–3 times the amount of K_2O that glauconite contains, hence contributes 2–3 times the amount of radiogenic argon for equal weights of the two minerals. This factor must be multiplied by approximately the true age of the orthoclase divided by the true age of the glauconite. It may be readily seen how a small percentage of contamination can lead to a 10–20 per cent error in the determined age.

Fortunately the contamination error can be assessed. The per cent volume contamination of a glauconite sample can be measured readily by the methods discussed above. A maximum reasonable error can then be assigned on the basis of the probable potassium content and the maximum reasonable age of the contaminating grains. Samples for which such a maximum reasonable error is excessive are best discarded. In case a date for such a sample is vital to the research at hand, a contamination correction can be arrived at from potassium and argon measurement on the contamination alone—that is on the fraction of the original sediment which most nearly approximates the contaminating material.

Argon Inheritance

At crystal formation, a certain amount of argon is incorporated into the structure. If the argon is of the same isotopic composition as present atmospheric argon, the A^{40} involved will be subtracted in the process of the atmospheric argon correction, and no error will result. If there is excess radiogenic argon in the initial argon of crystallization, however, the apparent age will be larger than the true age and the determination will be in error owing to radiogenic argon inheritance.

Lipson (1958) finds amounts of atmospheric argon in glauconite samples which are substantially in excess of air argon accidentally introduced during the analysis and which are noticeably greater than those in typical igneous samples. Thus his glauconite samples average 0.8×10^{-6} cc STP/gm atmospheric argon in excess of that found in blank runs whereas

typical igneous micas contain less than a third of this amount, and specimens of extrusive igneous mica have been encountered (Curtis, Lipson, and Evernden, 1956, p. 1360) in which the atmospheric argon content is a tenth of this amount. The question arises as to whether a significant fraction of this initial argon can be radiogenic in composition.

If the material crystallizes at the water-sediment interface, the inheritance error will not arise. Sea water has a large amount of dissolved argon in equilibrium with argon in the atmosphere. Any tendency for the local argon at the site of crystallization to be enriched in radiogenic A^{40} must be completely masked by the rapid exchange with dissolved argon from the atmosphere.

Argon inheritance could be a serious problem where the sedimentary mineral crystallizes after burial at the expense of decomposing detrital material rich in potassium and radiogenic argon. (Incomplete destruction of the primary lattice during growth of the secondary lattice could lead to especially serious examples of inheritance error.) The effect is diminished if the depth of burial is shallow and if the argon is mobile in the network of pores, fissures, and grain boundaries, providing contact between sea water and the site of crystallization. Although his data have not been substantiated, Pettersson (1953, p. 255) found evidence for the migration of radium in marine cores which seems to support the view that superfluous elements can migrate easily in newly deposited sediments, and thus the view that argon inheritance error in sediments should not be severe. Data on very recent sediments are needed to settle the question.

Diffusion

General statement.—All constituents of a crystal move about in the lattice by diffusion so that argon formed by K^{40} decay will to some extent migrate to crystal boundaries. Because of presumed increased mobility there, it is assumed that argon quickly escapes so that a boundary condition of zero concentration is closely approximated. Although the intergranular concentration of argon gas could become significant at depth and give rise to a reverse diffusion into the specimen grains, in the majority of environments, especially for sediments, the reverse diffusion should be negligible. Thus diffusion leads to an underestimate of age.

To correct for diffusion loss in a given specimen, the diffusion coefficient for argon in the specimen and the average grain size must be known. Unfortunately there are few data available concerning the diffusion coefficient for argon in crystals. At present there is no evidence that argon diffuses from micas. In feldspars, the effect can be appreciable as seen in the lower argon yields from feldspars than from contemporaneous micas; studies of this question have succeeded only in establishing the magnitude of the discrepancy in typical cases and are of little use in computing diffusion coefficients. The sole important attempt to treat argon diffusion quantitatively seems to be the work of Gentner and associates on Oligocene sylvites from Germany. These studies, which deal with alkali halides, may not be of central importance because K-A dating is more frequently applied to silicates, in which the mechanism of argon diffusion may be entirely different. Alkali-halide crystals, however, are among the best understood in solid-state theory, and it is in such crystals that we might hope first to understand the rate of argon diffusion quantitatively.

Gentner papers on argon diffusion in sylvite.—Diffusion was detected in lower Oligocene sylvite samples from Buggingen, Germany, and Alsace, France, by the observation that the radiogenic content of presumably contemporaneous salts varied systematically with the grain size of the samples. In the first of the papers in which the diffusion problem was examined, Gentner, Präg, and Smits (1953, p. 17) found that the experimental data for six salt samples, ranging in grain size from 0.5 to 4 mm, could be fitted on a single theoretical curve if the age of the deposits was taken to be 21 million years, and the diffusion constant for argon, D, had been constant throughout this lifetime at 1.5×10^{-19} cm^2/sec. Two other samples deviated severely from the pattern of the six, but there was evidence of crystal distortion in these samples.

In a later paper Gentner, Goebel, and Präg (1954, p. 124) restricted the study to Buggingen salts but measured the helium content as well as the argon content. From radioautographs of the samples, it was possible to measure the specific alpha activity, which is the present rate of helium production. This quantity varied from sample to sample (as much as 50 per cent in the extreme), but an average value was assumed to apply to all the salts studied. As anticipated, the helium content also varied with grain size, but less sharply than would be consistent with a constant value of D throughout the life of the specimens. Although it seems that the helium data are too low in precision and the sampling

too small in extent to justify such a step, these workers supposed that the data were adequate but that D for both argon and helium had declined exponentially with time. Such an exponential decline would result from a constant rate of cooling of the deposits. For their calculations, the writers supposed that the fractional decline in D is identical for the two gases. This implies that the heats of activation for argon and helium diffusion in the specimens are alike, a supposition which is perhaps easier to disbelieve than to accept. Nonetheless, values of the initial diffusion coefficient for argon, the initial diffusion coefficient for helium, the common ratio of initial to final diffusion coefficients, and the age of the deposits were then sought which would give best possible agreement with the observed argon and helium yields from five samples of varying grain size. The best fit corresponded to D (argon) declining from 5×10^{-18} to 5×10^{-20} cm^2/sec., D (helium) declining from 12×10^{-18} to 12×10^{-20} cm^2/sec., and the age of the deposits as 25 million years. The fit with the experimental data was about as good as it was in the earlier paper where constant D was assumed for argon; again several points deviated severely from the theoretical pattern.

In the most recent paper of the series, Gentner and Trendelenberg (1954, p. 261) applied an ingenious method for measuring the diffusion of gases in solids to a determination of the coefficient of diffusion for helium in sylvite. The method consists in heating the salt in a chamber communicating with the ion source of a continuously operated mass spectrometer of high sensitivity for helium. If the helium is initially distributed uniformly in the salt, and if an appreciable fraction of the total helium is driven out by preliminary heating, the gas flow rate will decline exponentially with time. From the characteristic time of the exponential and the grain size of the material, the diffusion constant at the temperature in question is easily computed. Diffusion rates in pure salts can also be determined if single crystals of the salt are crushed, sieved to a desired grain size, "tempered" by heating in a helium atmosphere until the diffusion process brings about a uniform helium content, and then studied by the same technique as described for the natural crystals.

Using this technique, precise values for the coefficient of diffusion of helium in pure NaCl were obtained for the temperature range 500°C. to 600°C., where the results can be represented by $D = 2.9\ e^{-13,700/T}$ cm^2/sec. Results for two natural sylvites were considerably less satisfactory. In both sylvites a low-temperature diffusion was found in the range 220°C. to 250°C. In one case (Wintershall sylvite) the helium loss in this range was so complete that no high-temperature diffusion was detectable. In the other sylvite (Buggingen) the low-temperature helium loss amounted to only ten per cent so that high-temperature measurements could be made. At about 435°C. a highly erratic helium evolution was observed which the authors attributed to escape of helium from inclusions fractured at this temperature, *i.e.* to decrepitation. A smooth helium evolution was observed again in the range 500°C.–650°C., but the characteristic times appeared to be independent of initial grain size. Gentner and Trendelenberg found that by assuming a post-decrepitation grain size of 0.4 mm for all samples, the calculated diffusion coefficients could be brought into fair agreement with those measured for NaCl. They then extrapolated D for helium in pure NaCl to 80°C. and 40°C. to get values of 40×10^{-18} and 32×10^{-20} cm^2/sec. respectively, which are not far from those inferred for the diffusion of helium in the Buggingen sylvites.

K^+ Diffusion in sylvite.—The interpretation of results described in the preceding paragraphs begins with an inquiry into the mechansim and the rate of diffusion of potassium ions in KCl. This process is thought to be fairly well understood. The question is treated in detail in a recent review by Seitz (1954, p. 11).

The K^+ diffuses in KCl by jumping into adjacent K^+ vacancies. At low temperatures the number of such vacancies per unit volume, n_+, is determined by properties (such as the percentage divalent-impurity content) of the individual specimen and is independent of temperature. Here the coefficient D for K^+ ions can adequately be represented by the expression

$$D = \nu_0 a^2 [\exp(-\epsilon_+/kT)] n_+/N \qquad (1)$$

where ν_0 is a frequency of the order of the lattice vibrations (10^{13} sec.$^{-1}$), a is the jump distance, ϵ_+ is the barrier or activation energy for the jump of a K^+ ion into an adjacent vacant site, and n_+/N is the constant fraction of such vacant sites in the lattice. The activation energy, ϵ_+, is about 0.85 electron volts so that the temperature dependence is exp $(-9855/T)$. At high temperatures, positive-ion vacancies and an equal number of anion vacancies (Schottky defects) are generated thermally by the transport of ion pairs from the interior of

the crystal to a surface. The fraction of positive-ion sites vacant when equilibrium with respect to this process occurs is

$$n_+/N = \exp(-\epsilon_f/2kT) \qquad (2)$$

where ϵ_f is the energy of formation of a vacancy pair. Thus as the temperature is increased, there occurs a "knee" in the plot of log D versus $1/T$ above which the temperature dependence is governed by the composite activation energy $\epsilon_+ + (\epsilon_f/2)$ which equals about 2.1 electron volts and corresponds to exp $(-24100/T)$. Since migration of K^+ ions is responsible for almost all the electrical conductivity in KCl, the diffusion coefficient need not always be measured but can often be inferred from electrical conductivity data through the Einstein relation

$$D = \frac{9 \times 10^{11} kT\sigma}{Ne^2}$$

where σ is the conductivity in ohm^{-1} cm^{-1} and e is the electronic charge in electrostatic units.

Since the knee in the log D and log σ versus $1/T$ curves occurs near 560°C. in the purest KCl crystals studied, the diffusion of K^+ at temperatures below this will always be impurity controlled and will vary from sample to sample. To compare a reasonable rate for K^+ diffusion at room temperature with the values found by Gentner *et al.* (1954, p. 130) for argon diffusion, suppose that the mole ratio of divalent impurity to potassium is 10^{-4} in a typical sylvite. These doubly charged impurities force an equivalent number of K^+ vacancies into the crystal to preserve electrical neutrality and make n_+/N equal to 10^{-4}. Taking ν_0 as 10^{13}, and a as 4.4×10^{-8} cm, one calculates from equation (1) for 40°C.

$$D(K^+) = 4 \times 10^{-20} \text{ cm}^2/\text{sec.}$$

which corresponds closely to what Gentner *et al.* (1954) have deduced for argon in KCl at this temperature. Thus there is no basis for the objection by Wasserburg and Hayden (1955, p. 59) that the room-temperature diffusion coefficient for Na^+ in NaCl differs by 12 orders of magnitude from the Gentner value for the argon diffusion coefficient. These writers have overlooked the fact that the high-temperature data for Na^+ diffusion cannot safely be extrapolated to low temperatures because of the knee in the log D *versus* $1/T$ plot.

Possible mechanism for argon diffusion in sylvite.—Considering now a subject which has been neglected in the literature, there are good reasons to expect that the mechanism of argon diffusion in sylvite differs from that of the diffusion of K^+ in sylvite. Because of the high ionization potential of argon, the rare-gas atom will diffuse as a neutral particle. The radius of this particle, as judged from the minimum separation in solid argon and from kinetic theory data, is large. That is, on the scale of conventional ionic radii, the argon atom would seem to be close to the Cl^- ion at 1.81×10^{-8} cm, whereas the K^+ ion is at 1.33×10^{-8} cm. The size of the argon atom, as it is needed for overlap-energy calculations, is important in estimating the activation energy for various kinds of jumps within the sylvite lattice. Moving as a neutral particle, the electrostatic and polarization terms are less prominent than in ion migration. The overlap or repulsive energy term dominates matters. If the radius of the chlorine-ion type for argon, estimated above, may legitimately be used for overlap-energy calculations (Mott and Gurney, 1948, p. 3) the following conclusions are readily established:

(1) At low temperatures, argon atoms will populate available Cl^- positions.

(2) The barrier energy for migration of the argon atom from one Cl^- position to an adjacent Cl^- vacancy is high enough (about 6 electron volts) so that this mechanism of diffusion is distinctly not available for argon.

(3) Because of its neutrality the argon atom can, however, move easily along the lines joining nearest neighbors in the KCl lattice and can thus diffuse by jumping through a positive-negative vacancy pair wherein the two vacancies are adjacent or associated. The activation energy for this jump is about 1 volt.

If this is the correct mechanism, the diffusion coefficient would depend on the density of associated vacancy pairs. At low temperatures, this density should be close to that for negative ion vacancies (assuming these are less abundant than positive ion vacancies) since the energy for dissociation of the pair is appreciable. Reitz and Gammel (1951, p. 898) have calculated this as about 0.9 electron volts in the case of NaCl. Furthermore the associated vacancy pairs are judged (Seitz 1954, p. 9) to be sufficiently mobile so that the jumping of the argon atoms continues to control the rate of diffusion. According to this picture, then, the low-temperature diffusion coefficient would depend on the density of potassium-chlorine vacancy pairs and should exhibit an activation energy of about 1 volt. It is difficult to estimate the density of vacancy pairs in a naturally occurring

alkali-halide crystal. Negative-ion vacancies will arise during crystal growth if divalent negative impurities occur in the crystal. But these should be rare in a crystal grown from solution since atomic divalent negative ions are uncommon in solution. In the absence of other mechanisms for generating negative ion vacancies, then, it would appear that the density of vacancy pairs should be substantially lower than the density of single positive-ion vacancies. This in turn would lead to substantially lower rates for the diffusion of argon atoms than for the diffusion of potassium ions.

Discussion of the Gentner results.—Results of the Gentner work which are precise are the values for the diffusion coefficient of helium in pure NaCl at elevated temperatures. As has been pointed out above, these extrapolate to a value of 3×10^{-19} cm²/sec. at 40°C., the present temperature of the Buggingen deposit. The mechanism by which the compact helium atom diffuses in alkali-halide crystals may be quite different from the mechanisms discussed above for diffusion of the K^+ ion and A atom. It is fairly certain, however, that impurities in the lattice enhance the diffusion rate rather than retard it so that the above extrapolation should give a lower limit for the diffusion coefficient at the low temperature. That Gentner *et al.* (1954, p. 130) arrive at a lower value of $D = 1.2 \times 10^{-19}$ appears as very strong evidence that their analysis regarding D as a function of time cannot be quantitatively correct. This analysis depends strongly on the helium data since the argon data alone can equally well be explained by a constant diffusion coefficient. Gentner and Trendelenberg (1954, p. 265) have proved in their work on pure NaCl, that helium is quite mobile in pure alkali halides. Helium mobility is probably even higher in geological samples and can be expected to vary appreciably with impurities from sample to sample. Indeed, Gentner and Trendelenberg found this in the case of the Wintershall sylvite where helium loss was complete at 250°C. The present authors believe that fine points, such as secular changes in the coefficient, concerning argon diffusion should be established with a more reliable rather than a less reliable auxiliary physical time scale.

The evidence for diffusion of argon to the extent of $D = 1.5 \times 10^{-19}$ cm²/sec. in the Buggingen and Alsace sylvites remains. Until there is more knowledge about the mechanism of argon diffusion and the activation energies, frequency factors, and vacancy densities involved, it is impossible to say more on theoretical grounds than that this value is reasonably close to those known for K^+ diffusion but that argon diffusion should perhaps proceed much more slowly. The Gentner measurement of this quantity may be criticized, however, since the assumption is made that the diffusion coefficient for all the samples is the same. This is not to be expected theoretically since the low-temperature diffusion coefficient must surely depend on the density of lattice vacancies of some sort, and this density will vary from sample to sample. A better method to determine D is to measure the argon yield from specimens for which an independent, reliable age is available.

Lipson's data on a Devonian sylvite.—Lipson (1958) (*see* also Folinsbee, Lipson, and Reynolds, 1956, p. 65) finds that the K-A date for a Devonian (Givetian) sylvite from Saskatchewan is 285 m.y. in agreement with the Holmes B time scale. The nearest calibration point on the Holmes scale is provided by a Connecticut Samarskite with concordant uranium ages of 255 million years (Nier, Thompson, and Murphey, 1941, p. 113). The maximum reasonable separation of this Samarskite and the Saskatchewan salt is half the Carboniferous period plus half the Devonian period. The Holmes estimates for these half periods would be 25 million years and 27 million years, respectively, totalling 52 million years. If this total is increased to 70 million years for safety a reasonable upper limit for the age of the sylvite is 325 million years.

The salt deposit has been described in detail by Cole (1948, p. 83). The average size of the sylvite grains in the specimen dated was 0.5 cm. The rich potassium salt bed from which this sample was taken (at a depth of 3472 feet) is thought to be an ideal bed for K-A dating as the dip of the beds in this area is only about 10 feet per mile, and the warping of the area has been very gentle. After deposition, the salt beds were buried under Paleozoic sediments to a possible depth of 2000 feet. Since that time the closest the salt came to the surface was 1200 feet so that the samples have been effectively protected from ground-water action. The estimated temperature at this level is 50°C., similar to the temperature of the Oligocene deposits in Germany. Assuming spherical grains of diameter 0.5 cm, the diffusion-corrected age of this salt was calculated for various values of the diffusion coefficient, D. Results for D equal to 0, 5×10^{-20}, 1.5×10^{-19}, and 5×10^{-19} cm²/sec. are 285, 346, 430, and 1250 m.y. respectively. From these and the upper limit of

325 m.y. for the age, D can be considered less than 3×10^{-20} cm^2/sec.

The difference between diffusion in the sylvite from Saskatchewan and those from Germany may be one of sample purity. The Oligocene potassium salts of Germany however, occur in a badly faulted and folded structure, the Rhine Graven. In a given bed, the coarse-grained salt might possibly be a part of the salt bed not affected by dynamic metamorphism, whereas the fine-grained salts might be recrystallized. Thus the coarse- and fine-grained crystals, even though in the same bed, might not be cogenetic.

The studies by Gentner and his associates of diffusion in Oligocene sylvites illustrate the kind of thorough measurement programs needed, but do not seem to be conclusive. Their results for helium diffusion are discordant with their precise values for diffusion of helium in pure NaCl at elevated temperatures. Without the helium results, there is no need to postulate time-diminishing diffusion at Buggingen. The value derived for the argon diffusion coefficient in these deposits needs to be confirmed by argon measurements on salts of known age from completely unfolded basins. Even so, variations from sample to sample may be anticipated. Lipson's data for a Saskatchewan sylvite indicate a lower value of D for this sample. Much further work is needed, on all important minerals involved, before the effects of diffusion on K-A dating can be gauged correctly.

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GEOLOGY OF THE MOUNT GARIBALDI MAP-AREA, SOUTHWESTERN BRITISH COLUMBIA, CANADA

PART I: IGNEOUS AND METAMORPHIC ROCKS

BY W. H. MATHEWS

ABSTRACT

The Mount Garibaldi map-area is in the southern part of the plutonic complex comprising the Coast Mountains of British Columbia.

The oldest rocks of the map-area are metavolcanic and minor metasedimentary rocks of unknown age and structure. These are extensively invaded by a quartz diorite batholith, the earliest of the plutonic masses, which displays an unfoliated core, a margin with highly developed secondary foliation, and an intervening zone of primary foliation. Bedded rocks at least 20,000 feet thick, principally clastic sediments derived in part from the batholithic rocks, occur in fault blocks north of Garibaldi Lake. The rocks of one fault block rest uncomformably on the older quartz diorites and contain Upper Cretaceous fossils; those of the other fault blocks are thought to be slightly younger than the fossil-bearing beds. The bedded rocks have in turn been invaded by later quartz diorites which are partly contemporaneous with and partly later than the block faulting. Two post-tectonic and presumably post-Upper Cretaceous batholiths lie partly within the map-area.

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INTRODUCTION

Location and Access

The Mount Garibaldi map-area, in the southern part of the Coast Mountains of British Columbia, covers 380 square miles between Lat. 49°38′ and 50°00′ N., and Long. 122°54′ and 123°14′ W. The town of Squamish, in the southwestern part of the map-area, lies at the head of Howe Sound, 41 miles by rail north of the city of Vancouver. The Pacific

Great Eastern railway extends northward past Squamish through the western part of the map-area, and a few gravel roads serve the southwestern part. Elsewhere access is limited almost entirely to foot trails.

Field Work

Most of the field work in the map-area was carried on during the summers of 1946 and 1947, although previous work had been done during brief visits since 1936. Information was plotted in the field on manuscript maps (scale 1:40,000, contour interval 100 feet) prepared by the British Columbia Department of Lands and Forests from photo-topographical surveys. Locations in the field were made by pace-and-compass traverses, by compass resections on identifiable landmarks, and by aneroid readings. Aerial photographs were limited to a few random obliques at the time of field mapping, but the entire area was covered by vertical aerial photography in 1949, and these photos have been used in the preparation and revision of this report.

Contacts between the batholithic rocks and older schists and sedimentary rocks were crossed transversely at regular intervals, generally along creeks or ridge crests; but as far as was practicable most of the contacts of the older rocks with the overlying Quaternary volcanic rocks were traced directly.

Previous Work

Reference has been made in the literature to the geology of parts of the area by O. E. LeRoy (1908), C. Camsell (1918), and F. C. Bell *et al.* (1932). E. M. J. Burwash included reconnaissance observations from the Garibaldi Lake area in his doctorate dissertation (1918). H. T. James (1929) has made a detailed survey of the Britannia Beach map-area which adjoins the Mount Garibaldi map-area on the south. Full descriptions of mining properties on Ray and Raffuse Creeks are given in the 1937 annual report of the Minister of Mines of British Columbia by B. T. O'Grady (1938), and briefer descriptions of these and other properties of the map-area are given in earlier annual reports of the same series.

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Geology of the Igneous and Metamorphic Rocks

Metavolcanic and Metasedimentary Rocks

Rocks mapped as metavolcanic and metasedimentary rocks are exposed over an area of approximately 27 miles within the map-area, principally near its southern end. They include such varied rock types as massive and bedded greenstone, chert, argillite, sandstone, conglomerate, and the schistose equivalents of these rocks. Crystalline limestone occurs as lenses generally only a few feet wide in the northernmost part of the map-area. The stratigraphy has not been worked out.

In the relatively few localities in which stratification has been observed (Fig. 1) it displays a general northerly strike but highly variable dips, changing markedly in the incompetent argillites within an area of a few square feet. Faults with displacements of tens of feet or more are common and make it impossible to establish the sequence of beds. Although only two faults with mappable displacements have been recognized in the field—one west of Mount Baldwin (Pl. 1), the other at the south edge of the map-area near the head of Raffuse Creek—it is possible that the irregular outline of the areas of metavolcanic and metasedimentary rocks elsewhere may be attributed in part to large-scale faulting.

Most of the rocks of this unit have undergone dynamic metamorphism with the development of chlorite, but not biotite. Schistosity is most marked in the vicinity of the quartz diorite intrusive rocks, and two areas remote from these intrusive rocks—one northeast of Daisy Lake, the other on Goat Ridge west of Sky Pilot Mountain—are notably unsheared. A considerable area at the latter locality is, moveover, marked by nearly horizontal bedding.

At least a part of this map unit can be correlated with the contiguous Goat Mountain

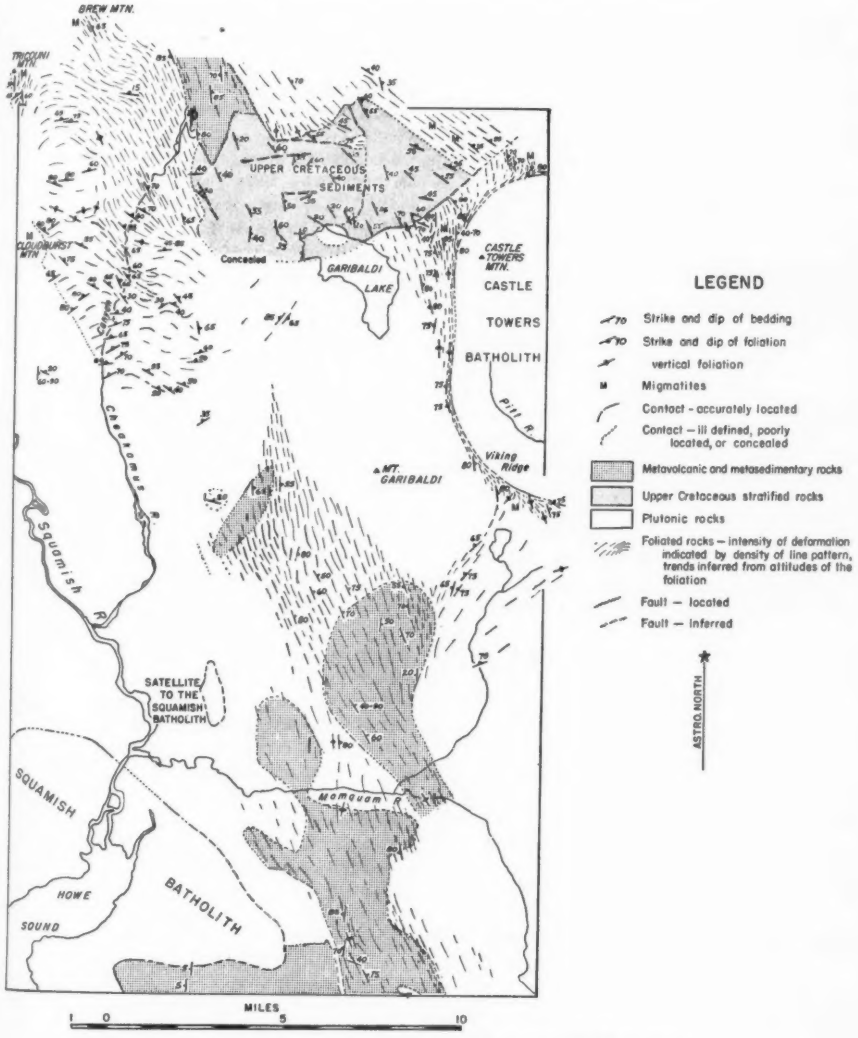

FIGURE 1.—TECTONIC MAP OF THE MOUNT GARIBALDI MAP-AREA

Formation of the Britannia Beach map-area (James, 1929, p. 15–24), but it has not been possible to recognize in the Mount Garibaldi map-area the three subdivisions of this formation.

The northernmost area of metavolcanic rocks, which differs from those farther south in containing lenses of limestone and in lacking clastic beds, may be a distinct stratigraphic unit of a different age.

No definite information on the age, or ages, of these rocks has been found, other than that they antedate the Cloudburst quartz diorites which are themselves pre-Upper Cretaceous. Successions of metavolcanic and metasedimentary rocks are common within southwestern British Columbia (Mathews, 1947, p. 34–37) in the upper Paleozoic, the Triassic, and the Jurassic, and the rocks of the Mount Garibaldi map-area may correspond to one or more of these.

TABLE 1.—STRATIGRAPHY

Age	Formation or Map Unit	Lithologic Characters	Thickness (feet)
Quaternary		Alluvium, slide debris, lake beds, moraine, outwash	Zero to several hundred
		Terraced gravels and fanglomerates, raised marine clays and silts	Zero to several hundred
	Garibaldi Group	Andesite and basalt flows and cinder cone; dacite flows, glowing avalanche deposits, and intrusives; rhyodacite intrusive	Zero to several thousand
		Erosional unconformity	
Probably Post-Upper Cretaceous	Castle Towers batholith	Quartz diorite, trondhjemite, minor pegmatitic and other dikes (post-tectonic)	.....
		Relations unknown	
	Squamish batholith	Granodiorite, quartz monzonite, minor pegmatitic and other dikes (post-tectonic)	.....
		Relations unknown	
Post-mid-Upper Cretaceous	Younger quarts diorites	Quartz diorite stock, dikes, sills, major plutonic rock (syntectonic and post-tectonic)	.....
		Intrusive contact	
Probably mid-Upper Cretaceous or later	Helm Formation		
	fifth member	Paragneiss	Several thousand (?)
	fourth member	Principally slaty argillite, minor quartzite	3500±
	third member	Conglomerate, quartzite, calcareous sandstone, sandy limestone, slaty argillite	2000±
	second member	Slaty argillite, minor argillaceous quartzite	3250±
	first member	Argillaceous quartzite, greenish arkose, conglomerate, argillite	2000±
		Fault contact	
	Empetrum Formation	Metavolcanic rocks, limestone, conglomerate, minor grit and argillite	Several thousand
		Fault contact	
Mid-Upper Cretaceous	Cheakamus Formation		
	upper	Greenish conglomerate, graywacke	3000±
	middle	Stratified graywacke, arkose, argillaceous and calcareous beds, locally fossiliferous	1000±
		Massive green graywacke	5000±
	lower	Greenish and grayish conglomerate, arkose, argillite.	Few hundred to several thousand (?)
		Erosional unconformity	
Pre-Upper Cretaceous	Cloudburst quartz diorites	Major pluton or plutons of quartz diorite, diorite; quartz feldspar porphyry; orthogneiss, augen gneiss, mylonite. green and white schists	.....
		Intrusive contact	
	Metavolcanic and metasedimentary rocks	Massive greenstone, breccia, conglomerate, slaty argillite, limestone; green schist	(?)
		Base concealed	

Plutonic Rocks—Part 1

General Statement.—Granitic rocks underlie the greater part of the map-area and can be subdivided into at least four map units. The largest of these, a heterogeneous assemblage of foliated and unfoliated quartz diorites[1] and diorites, is clearly overlain unconformably by the mid-Upper Cretaceous Cheakamus formation and is referred to here as the Cloudburst quartz diorites. Several bodies intrusive into the Upper Cretaceous beds will be referred to as younger quartz diorites. The western segment of another large pluton made up of quartz diorite and trondhjemite in the northeastern corner of the map-area will be referred to as the Castle Towers batholith. Finally, a mass of quartz-rich granodiorite and quartz monzonite in the southwestern part of the map-area will be referred to as the Squamish batholith and a smaller body of the same rocks near-by as its satellite stock. Although neither the Castle Towers batholith nor the Squamish batholith and its satellite are in contact with the Upper Cretaceous beds, all have escaped the extensive deformation which has tilted and block faulted the stratified rocks and the plutonic rocks are, therefore, considered to be younger than mid-Upper Cretaceous.

Most if not all of the plutonic masses of the map-area can be grouped within the Coast intrusions, to use the term now adopted by the Geological Survey of Canada in place of the misleading title Coast Range batholith for the great plutonic complex exposed in and adjacent to the Coast Mountains of British Columbia, Alaska, and Yukon Territory. Should the term Coast intrusions be restricted to pre-Tertiary plutonic rocks as proposed by Rice (1947, p. 33) in his original definition, or more specifically to syntectonic or pretectonic intrusions as suggested by Aitken (1955, p. 1701), the Castle Towers and Squamish batholiths would not be included nor would the quartz diorite stock on the north shore of Garibaldi Lake.

The diversity in character and age of the plutonic rocks of the map-area conforms to the observations of Cairnes (1924, p. 82–108), Dolmage (1929, p. 85–87), Duffel and McTaggart (1952, p. 78–83), and a few others regarding the composite nature and extended age range of the granitic masses in the western part of the North American cordillera.

Cloudburst quartz diorites.—The cloudburst quartz diorites make up almost two-thirds of the exposed basement rocks. Two plutons, separated by a discontinuous belt of metavolcanic and metasedimentary rocks (Fig. 1) may be present, but it has not been possible to recognize significant differences between the rocks on either side of the belt of metamorphic rocks, and all may be the product of a single intrusion.

TABLE 2.—SURFACE AREAS OF MAP UNITS
(Total area mapped—370 square miles)

Map Unit	Percentage at Present Surface		Percentage at Pre-Garibaldi surface	
Lakes, Howe Sound	3.6			
Glaciers	6.1			
Moraines	1.0			
Alluvium, slide debris	6.4			
Terrace gravels, fanglomerate	1.7			
		18.8		
Garibaldi group rocks		15.7		
Younger plutonic rocks				
Castle Towers batholith	3.5		5.4	
Squamish batholith	7.6		11.6	
Younger quartz diorites	0.3		0.5	
		11.4		17.5
Stratified rocks of the Garibaldi Lake area				
Helm Formation	1.1			
Empetrum Formation	0.2			
Cheakamus Formation	2.9			
		4.2		6.4
Older plutonic rocks				
Cloudburst quartz diorites		40.9		62.4
Metavolcanic and metasedimentary rocks		7.4		11.3
Migmatites of various ages		1.6		2.4

Quartz diorite predominates in these older plutonic rocks, but locally more basic rocks, such as the diorite of Cheakamus Canyon, or more siliceous rocks, such as the quartz-feldspar-muscovite schist of Watson siding, are important. Potassium feldspar is absent or nearly so in all rocks of this unit. In some areas later stocks or dikes may be included, but all have apparently had a similar tectonic history.

Three mutually gradational structural types can be recognized: (1) unfoliated quartz diorite, (2) quartz diorite and diorite with a primary

[1] Lindgren's divisions are used in the classification of the granitic rocks.

foliation, and (3) quartz diorite with secondary foliation locally converted to mylonite. Migmatites of limited areal extent also occur.

UNFOLIATED QUARTZ DIORITE: Unfoliated and relatively unaltered quartz diorites occur on both sides of Squamish Valley above the mouth of Cheakamus River. They are characterized either by biotite and hornblende or by hornblende alone. No systematic distribution of these two types is evident. The oligoclase or andesine commonly displays incipient saussuritization. Similar fresh to slightly altered quartz diorites are present in upper Mamquam valley, northwest, north, and east of Mount Garibaldi and along the lower slopes of Squamish Valley below Cheakamus River.

PRIMARILY FOLIATED QUARTZ DIORITES AND DIORITES: Quartz diorites and diorites exhibiting primary foliation are well displayed in Cheakamus Canyon for about 2½ miles along the Pacific Great Eastern Railway south of the Cheakamus bridge, and in adjacent areas east for almost 3 miles and northwest beyond the limit of the map-area. Strong preferred orientation of hornblende and biotite is obvious in nearly all outcrops, and locally dark elongated basic clots parallel this foliation. Planar structure is further emphasized in most thin sections by a strong preferred orientation of the subidiomorphic plagioclase crystals. The attitude of the planar structure is rarely uniform for more than a few hundred to a few thousand feet (Fig.1). Vertical air photos show on many of the outcrops a fine-textured lineation paralleling the observed strike of the foliation. Abrupt changes in the trends of this lineation occur at well-defined fracture traces. These sharp changes in attitude and equally abrupt changes in rock texture and composition can best be attributed to widespread block faulting.

Although the foliation here is ascribed to movements of magma before complete consolidation, all rocks studied in thin section have undergone some later deformation. Plagioclase grains commonly show undulatory extinction, and their twin lamellae may be visibly bowed; biotite flakes are also commonly bent, and locally quartz is granulated and strung out along microscopic shear planes. At least some epidote, which constitutes up to 8 per cent of the rock, may have developed during alteration of andesine to the oligoclase which now predominates. Biotite is commonly green, and this mineral and to a less extent the hornblende are partly converted to chlorite. The incomplete development of a new mineral assemblage, sodic feldspar-epidote-chlorite-quartz, from what was presumably the original assemblage, andesine - hornblende - biotite - quartz, marks a response to a low grade of dynamic metamorphism, tending toward the greenschist facies. Thin sections suggest that this postconsolidation deformation became more intense toward the northeast in the Cheakamus Canyon area.

SECONDARILY FOLIATED QUARTZ DIORITES: Secondarily foliated rocks have been studied in most detail in a northwestward-trending belt up to 4 miles wide north and east of Cloudburst Mountain where they are made up of sheared quartz diorites, augen gneisses, and mylonite. Probably no part of the quartz diorite of this belt has escaped some deformation. Intensity of shearing varies markedly within even a few feet, but it tends to increase toward the northeast.

The least sheared of the quartz diorites in this belt closely resemble the adjacent primarily foliated rocks of the Cheakamus Canyon, but deformation and alteration has converted original mafic minerals to ragged patches of chlorite. Where deformation has been more intense the rock is an augen gneiss consisting of ellipsoidal to nearly spherical eyes of oligoclase in a fine-grained light to dark-gray matrix consisting of granulated albite, quartz, muscovite, chlorite, and accessory minerals. The feldspar augen have neither the well-defined crystal outline nor the strong orientation of the plagioclase of the primarily foliated rocks. Chlorite occurs as minute oriented flakes scattered through the groundmass or concentrated into long subparallel or swirling trains, which, alone, render the foliation conspicuous in many of the hand specimens. Microscopic examination reveals similar trains of well-oriented muscovite flakes and of randomly oriented quartz grains. Accessory minerals include sphene and apatite; epidote is not abundant; strongly zoned crystals of allanite occur in one thin section.

Banded augen gneisses, consisting of streaks from 1 to 3 cm wide of chlorite-rich and chlorite-poor quartz diorite, occur in a few places. Neither field nor microscopic evidence indicates whether this banding has developed by some process of segregation during deformation or has been inherited from the original rock. Should the direction of shearing parallel original banding, such banding could be retained, but should the two directions intersect the later movement might destroy the earlier foliation. Chance parallelism of the two directions may

explain the apparently irregular distribution of the banded augen gneiss.

The structure and mineral assemblage of these secondarily foliated rocks indicate intense low-grade dynamic metamorphism in which original quartz diorites, and perhaps migmatites, have been converted to rocks of the muscovite-chlorite subfacies of Eskola's greenschies facies (Turner, 1948, p. 96). The same mineralogical alterations noted in their incipient stages in the orthogneisses of Cheakamus Canyon are here essentially complete.

Sheared quartz diorites also occur in the southern and central parts of the map-area on either side of and along the extension of a belt of older metavolcanic and metasedimentary rocks. The quartz diorites are here so strongly sheared, and chlorite is so widely developed that many of the resulting rocks cannot be distinguished in the field from metavolcanic rocks, and only their general uniformity and the absence of interbedded sedimentary rocks suggest a plutonic origin. Thin sections of these rocks reveal lenses of granulated quartz and eyes of albite or oligoclase such as could form from quartz diorites, or quartz-feldspar porphyries. Inasmuch as quartz porphyries are not known in the unsheared metavolcanic rocks except as intrusive masses, the inclusion of these schistose green augen gneisses with the plutonic rocks seems justified. Precise delineation of the contact between green schists of plutonic origin and those of volcanic origin, however, remains far from satisfactory.

MIGMATITES: Migmatites are present in a few localities within both the foliated and unfoliated quartz diorites. The higher part of Cloudburst Mountain is an intrusion breccia consisting of migmatite in numerous angular unoriented blocks, from a fraction of an inch to scores of feet across, frozen in a matrix of unfoliated quartz diorite. The belts of orthogneiss and secondarily foliated quartz diorite extending across Cheakamus Valley are here truncated by the unfoliated quartz diorite, and possibly many of the banded migmatites have been developed from the foliated rocks. The relationships on Cloudburst Mountain are repeated on the southwestern slope of Tricouni Mountain, 1 mile north of the northwest corner of the map area (Fig. 1). Here, however, some amphibolite is included in the migmatite and possibly some of the metamorphic roof rocks may be present. Migmatites also occur on Brew Mountain, 2 miles northeast of Tricouni Mountain, but here the relationships to younger intrusive rocks are unknown. Banded migmatites are common in the Cloudburst quartz diorites of the northern, western, and southwestern slopes of Castle Towers Mountain where they occur as bodies up to a mile long and a third of a mile wide. Both the migmatite bodies themselves and the banding within them have a northerly to northwesterly trend. Migmatites also occur on the south face of Viking Ridge near the southwestern contact of the Castle Towers batholith. Many of the occurrences may be relics of the former roof and wall rocks of the plutonic rock; others such as that of Cloudburst Mountain, may be the foliated part of the Cloudburst quartz diorite itself migmatized during later intrusion of the unfoliated part. Other migmatites within the belt of secondarily foliated quartz diorites may have been partly disguised by shearing.

STRUCTURE OF THE CLOUDBURST QUARTZ DIORITES: The structure of the Cloudburst quartz diorite intrusive rock, though complex, permits some generalizations. The quartz diorite is intensely sheared almost everywhere along its contact with older rocks (Fig. 1). Such shearing extends locally as much as 4 miles into the plutonic rock. The older bedded rocks are also sheared but only within a few hundred to a few thousand feet from the quartz diorites. The sequence of rocks from wall to core of the intrusive rock, best displayed along Cheakamus Valley, consists of (1) unsheared greenstones and other constituents of the wall rocks, (2) green schist derivatives of the wall rocks, (3) schistose quartz diorites, augen gneisses, and mylonites developed from the intrusive rock itself, (4) primarily foliated quartz diorite and diorite, and (5) unfoliated quartz diorite. Here a general gradation exists between the first two and last three, but the contact between wall rocks and quartz diorites is concealed. Elsewhere the succession is similar except that primarily foliated rocks have not been detected, and in two localities the unfoliated quartz diorites are in direct contact with secondarily foliated quartz diorites along a zone of intrusion breccia. In some places belts of primarily foliated rocks may have escaped detection. The contact between wall rocks and quartz diorite is commonly disguised by the intense shearing.

The apparent gradation between unfoliated and foliated quartz diorites suggests that all were originally part of a single body of quartz diorite magma. Deformation affecting the wall rocks and outer part of the intrusion had no influence on the core but did give rise locally

to primary foliation in an intermediate zone. Movements may have taken place when the outer part of the intrusive rock was completely solid, the intermediate zone partly consolidated, and the core still molten. Later, a part of the molten core may have invaded the sheared margins to give rise to the internal disconformities, the intrusion breccias of Cloudburst and Tricouni Mountains and of the southwestern base of Brohm Ridge, and possibly the migmatites of Cloudburst Mountain. Then movement ceased, and the central portion consolidated without development of flow structures.

The localization of shearing around the margin of the pluton, the variations in its trend, and the absence of foliation in some of the near-by wall rocks suggest that deformation originated locally rather than from regional tectonic movements. These characteristics are like those of the Colville batholith of Washington, with its protoclastic border, which Waters and Krauskopf (1941) attribute to stresses originating in the upward movement of the batholith during emplacement. Three distinctions between the two plutons, however, are noteworthy: (1) foliation in the Cloudburst quartz diorites of the Mount Garibaldi map-area is almost invariably steeply dipping, whereas in the Colville batholith it describes broad domes and arches with dips only locally exceeding 30°, (2) foliation in the Mount Garibaldi map-area locally cuts across the contact between intrusive and older rocks, whereas in the Colville batholith it almost invariably parallels this contact, and (3) no zone of swirled and porphyritic gneisses is developed in the Cloudburst quartz diorites as it is in the Colville batholith. Notwithstanding these distinctions the two plutonic rocks may have had a similar history.

A possible objection to the hypothesis of deformation during, and by, intrusion lies in the low grade of metamorphism in the sheared rocks. Even in the orthogneisses showing incipient cataclasis biotite had become an unstable mineral before movement ceased, and at the time these rocks were in the lowest or chlorite zone of dynamic metamorphism where temperatures must have been relatively low. Such a condition is at least unexpected in the broad marginal portion of a batholith presumed to have a still-molten core not more than 2 miles distant.

AGE AND CORRELATION: The Cloudburst quartz diorites cut metavolcanic and metasedimentary rocks of uncertain age and are overlain unconformably by Upper Cretaceous sedimentary rocks. The time of plutonic intrusion in the southern Coast Mountains is now known to extend back to the Middle or Lower Jurassic (Duffel and McTaggart, 1952, p. 79), and in the absence of evidence of a still earlier age, the Cloudburst quartz diorites may tentatively be referred to the Jurassic or Lower Cretaceous.

If the cataclastically deformed marginal phases of the Cloudburst quartz diorites are to be regarded as a product of autometamorphism rather than an expression of external tectonic forces, there can be little other than age determinations by radioactivity on which to base correlations with other plutonic rocks in the Coast Mountains. In the "southwestern area of batholic rocks" of the Britannia Beach map-area, however, James (1929, p. 45–49) finds a sequence of primary and secondarily foliated granitic rocks which closely parallels that in the Cheakamus Valley. He states:

> "the four specimens show that the nature of the granitic rocks varies as the contact is approached and that the variations are of a uniform and progressive type. Within the line of section all the rocks are foliated; but at 3,000 feet from the contact, the parallel orientation of the grains was brought about during the crystallization period of the magma and is, essentially, a primary structure. As the contact is approached, this primary foliation is obliterated by a later shearing, which is parallel to the primary structure, and which becomes more intense until the rock, at the very contact, is completely recrystallized."

He notes, too, the presence of sodic plagioclase in the sheared rocks and the development in a few localities of chlorite and sericite. These rocks of the Britannia Beach map-area had a similar geologic history to that of the Cloudburst quartz diorites and may be of the same general age.

Some of the younger quartz diorites at the northern edge of the map-area are also sheared and closely resemble some of the Cloudburst quartz diorites although clearly of a different age. The shearing of these younger rocks was probably accomplished by external forces which also tilted and faulted the adjacent Upper Cretaceous sediments, and not by autometamorphism. If so, the core of the Cloudburst quartz diorites has escaped this later deformation, and the record of such deformations cannot be used as reliable age criterion.

Stratified Rocks near Garibaldi Lake

General statement.—Stratified rocks, chiefly clastic sedimentary rocks, underlie most of the map-area north and northwest of Garibaldi

Lake. These are subdivided into three map units, here called the Cheakamus, Empetrum, and Helm formations, which appear to be associated with one another in time and in circumstances of origin as well as in present spatial distribution. All three formations include conglomerate beds containing abundant granitic fragments, hence all three postdate the first period of batholithic intrusion in or near the map area. All three formations are cut by later granitic rocks. Only the Cheakamus Formation has to date yielded fossils, and the range of time over which these rocks were being deposited remains unknown. The middle part, and presumably all, of the Cheakamus Formation is of mid-Upper Cretaceous age. This formation rests directly on the Cloudburst quartz diorites and is probably the oldest of the three units. The Empetrum and Helm formations may have been deposited later in the Upper Cretaceous.

Interpretation of the stratigraphic succession of these rocks is made difficult by the almost complete absence of well-defined horizon markers as well as by structural complexity. Few beds are sufficiently distinctive to be traced in the field even in areas of good exposure, and these beds can rarely be followed for more than a few hundred feet before they terminate abruptly, apparently against transverse faults. Most of the major subdivisions of the formations, however, have a much greater continuity, and probably most transverse faults are small. A few larger discontinuities have been recognized in the field, and several have been identified as either a single major fault or a composite pattern of small faults. Conceivably other major discontinuities, notably those between the three formations, may be unconformities with marked overlaps.

Changes in lithologic facies may also contribute to difficulties in mapping, although few such changes can be demonstrated.

Cheakamus Formation.—The Cheakamus Formation, exposed on the eastern slopes of Cheakamus Valley, after which it is named, consists mainly of green arenaceous and conglomeratic beds having a general easterly dip.

The base of the formation is exposed at one locality, 1½ miles southeast of Garibaldi station, where massive conglomerate, made up of boulders of granitic rocks and of greenstone up to 2½ feet in diameter, rests directly on highly sheared Cloudburst quartz diorite. Higher parts of the succession are not exposed in this immediate locality, but similar conglomerates, interbedded with coarse sandstone, a graywacke, and minor gray to black argillite, are found 1 mile to the north on the lower slopes of Rubble Creek valley and again on the hillside southeast of Daisy Lake. Both grayish and greenish tints prevail in the coarser beds. Many of the rocks, especially those with a grayish argillaceous matrix, are highly sheared. Bedding is inconspicuous, and the few observations indicate widely varying attitudes. In view of the relatively narrow width of this belt of rocks, however, probably not more than a few hundred feet of beds are present.

The middle part of the Cheakamus Formation is an exceedingly uniform and massive green graywacke in which stratification can only very rarely be detected. Much of this graywacke might be confused in hand specimen with greenstone but can be distinguished from the latter rock with the aid of the hand lens by the abundance of minute quartz grains. Pebbles, generally less than 1 inch in diameter occur at a few localities either singly or clustered in small lenses. Observations on a few of these pebble lenses and on the rare and obscure stratification indicates that this member has a rather consistent easterly dip of 45°. This, and the observed outcrop width of approximately 1 mile, suggests a stratigraphic thickness of at least 5000 feet. The massive graywacke is overlain on the east by about 1000 feet of similar greenish sandstones in which stratification is revealed by an alternation of beds pale green noncalcareous and darker green with calcareous cement commonly from 20 to 50 feet thick, as well as by local color bands a few inches thick. A few strata are rich in pebble bands or shell fragments or both. Brown-weathering black argillaceous beds are locally present, and a bed of dark-gray coquina about 20 feet thick occurs near the base of the West Bluff of The Black Tusk. Well-developed cross bedding and less common ripple marks exist in some of the sandstones. What is probably a single horizon extending north and northeast from The Barrier to the base of the West Bluff is marked by several fossiliferous outcrops yielding fragmentary to nearly complete shells of large *Inocerami*.

Thin sections of graywackes from the middle members of the Cheakamus Formation show that detrital grains of quartz and plagioclase predominate in the rock. Fragments of mylonite, like that in the secondarily foliated Cloudburst quartz diorites only a few miles to the west are common. Grains of epidote, amphibole, sphene, and black opaque minerals are also present. Grains of allanite, like that in one specimen of the Cloudburst quartz diorites, (13) were seen in one thin section. The source of

scattered microcline grains on two thin sections is unknown. The matrix of the graywackes is commonly chlorite, and it is this mineral which contributes the green color to so much of the rock. Sericite forms the matrix of some of the lighter-colored beds. Cubes of pyrite are locally present.

The uppermost part of the Cheakamus Formation, exposed on the higher slopes south and east of The Black Tusk and on the northwestern slopes of Panorama Ridge, consists of an alternation of greenish coarse pebble conglomerate and distinctly stratified greenish graywacke. Bedding dips eastward or northeastward at from 10° to 55°, and as the width of this belt of rocks reaches 1½ miles a stratigraphic thickness of at least 3000 feet is indicated. A massive sandstone member exposed on Garibaldi and Lesser Lakes and on the bench half a mile to the north is not identifiable west of The Black Tusk.

A succession of grayish and greenish conglomerate and graywacke and lesser amounts of white to flesh-colored arkose and gray to black argillite is exposed in the northern part of the area mapped as Cheakamus Formation. These rocks are separated from the sequence already described by an easterward-trending fault, and their stratigraphic position remains in doubt. Attitudes on bedding are highly variable but commonly high, and judging from the 1½- to 2-mile outcrop width, several thousand feet of sedimentary rocks may be present. Lithologically they most closely resemble the lowermost few hundred feet of the Cheakamus Formation, as typically developed near Rubble Creek and Daisy Lake, but the apparent differences in thickness make this tentative correlation far from certain.

Empetrum Formation.—The Empetrum Formation is confined to a triangular area 2 miles long and 1 mile wide on Empetrum Ridge, after which the unit is named, and on the neighboring ridge to the west. Much of the formation is made up of greenstone and greenstone breccia derived in part from amygdaloidal and porphyritic lavas and in part from tuffs or graywackes. Lenses of white to cream-colored limestone, commonly containing lime silicate minerals and minor cream-colored dolomite, are present throughout the succession. Conglomerates containing abundant partly rounded granitic pebbles and boulders occur at several localities near the northern end of Empetrum Ridge. Argillite and coarse gray sandstone have been observed at one place in this same area. Much of the succession is highly sheared: schists are common, and pebbles in the conglomerate and markedly distorted. The elongation of the pebbles, a linear schistosity, and the axes of small folds plunge 30–40° to the southeast. No definite sequence can be recognized, and repetition by folding, if not by faulting, is probable. The width of the formation measured perpendicular to the strike is at least 1 mile and dips average 45° to the northeast; up to several thousand feet of beds may be present.

Helm Formation.—The Helm Formation is so named from its exposures on the southwestern, southern, and eastern margins of Helm Glacier. This formation underlies a total area of about 6 square miles of which about 25 per cent is concealed beneath ice or Quaternary volcanic rocks. The succession of sediments from Panorama Ridge on the southwest to the slopes above Cheakamus Lake on the northeast can be subdivided into 5 more or less distinct members.

The lowest member, fully exposed on the south face and west end of Panorama Ridge, is made up almost exclusively of arenaceous sedimentary rocks: bluish-gray argillaceous quartzites and less common pale greenish graywackes predominate; conglomerates, containing granitic fragments, and argillites occur in a few widely scattered beds. A thickness of 2000 feet has been measured on the south face of the ridge. On the north face, where the contact with the Cheakamus Formation truncates the bedding, only the uppermost few hundred feet of this member are present; still farther north the member is absent.

The second member, present but rather poorly exposed on and near the eastern end of Panorama Ridge, consists of massive argillaceous quartzites and slaty argillites. The latter rock is subordinate to the quartzites in the lower part of the member but predominates in the upper part. Stratification is exceedingly obscure throughout most of the succession, and the few observations on argillaceous beds indicate not only that these are highly and irregularly contorted, but also that they bear no constant relation in attitude to slaty cleavage. Three attitudes on stratification in quartzite are more consistent and indicate an easterly dip averaging about 45°. This and the outcrop width of about 0.9 miles, indicate a stratigraphic thickness of about 3250 feet.

Most of the third member is concealed by the eastern part of Helm Glacier, but limited exposures occur near both source and terminus of this body of ice. At the base, near the east tongue of Helm Glacier, a zone of conglomerate

and quartzite rests with a sharp but apparently conformable contact on the underlying argillites of the second member; the corresponding beds 1½ miles southeast are quartzites and calcareous sandstones. The middle part of the succession consists of coarse sandstone characterized by grains of both clear and bluish-opalescent quartz and by abundant dark minerals, overlain by pale-grayish-brown calcareous sandstone and sandy limestone. A few beds of slaty argillite are present. In the upper part of the member, made up chiefly of dark argillaceous quartzite, slaty argillite is conspicuous, and at the contact with the overlying fourth member it becomes predominant. What appears to be a complete and relatively undisturbed succession of beds of the third member, half a mile to 1 mile northeast of Gentian Peak, is about 2000 feet thick.

The fourth member consists predominantly of stratified argillite with minor quartzite. Bedding dips rather regularly at from 35° to 55° to the northeast and north across the full outcrop width, approximately 1 mile. The stratigraphic thickness is hence about 3500 feet.

Rocks still higher in the succession, exposed north and northeast of Corrie Peak, have been dynamically metamorphosed, and their original character is uncertain. They now consist of amphibole- and amphibole-biotite gneisses containing abundant quartz and plagioclase; the plagioclase occurs in some places as oligoclase porphyroblasts. Chlorite, epidote, sphene, pyrite, and garnet are locally present. Both amphibole and biotite exhibit a strong preferred orientation which with the larger-scale banding dip about 45° northeastward and parallel the bedding in the slates to the southwest. If the banding in these altered rocks corresponds to bedding in original silica-rich sedimentary rocks a stratigraphic thickness of at least several thousand feet may be present, and the upper limit has not been determined.

Metamorphism.—The general metamorphism of the Cheakamus Formation has been barely sufficient to recrystallize the chlorite occurring in interstices of the graywackes, conglomerates, and arkoses.

No significant metamorphism has been noted in the lower part of the Helm Formation on Panorama Ridge, but the higher part of this formation, southeast of Helm Peak, is distinctly metamorphosed, and the uppermost part, northeast of Corrie Peak, is extensively so. Although the gneisses of the latter locality show strong preferred orientation of mafic minerals, indicative of stress, the argillaceous rocks southeast of Helm Peak contain unoriented prisms of andalusite indicative of thermal rather than dynamic metamorphism. Garnet locally present in the migmatites suggests a higher grade of metamorphism than has been recognized elsewhere in the map-area.

The Empetrum Formation is distinctly sheared, and the effects of deformation are noticeable in the squeezing of the pebbles in the conglomerates. The development of lime-silicate minerals and the preservation of dolomite provide some limits to the grade of metamorphism.

Internal relationships.—The contact between the Cheakamus and Helm formations is concealed by moraine over much of its length and by a small body of ice on the north wall of Panorama Ridge. On the crest of the ridge arenaceous beds near the base of the Helm Formation are in contact with beds high in the middle member of the Cheakamus Formation. Half a mile to the north argillaceous beds 2000 feet higher in the Helm Formation adjoin conglomeratic beds of the Cheakamus Formation also about 2000 feet higher in the section than at the first locality. The stratification in the two formations is nearly parallel. If the contact between the formations is depositional, a striking disconformity exists. A fault relationship is more probable, but the existence and attitude of the inferred fault has not been established.

The Empetrum Formation is separated from the upper part of the Cheakamus Formation by later intrusive rocks and by a cover of moraine and Quaternary volcanic rocks, and from the middle part of the Helm Formation by either an angular unconformity, or more probably by an extension of the fault separating the Helm and Cheakamus formations. If the faulted relationship applies, the Empetrum Formation can best be regarded as intermediate in age between the Helm and Cheakamus formations and eliminated by the fault everywhere south of Empetrum Ridge.

Age and correlation.—The age of the Cheakamus Formation has been established by the *Inocerami* obtained from the upper part of the middle member of this formation at two localities, one immediately northwest of The Barrier, the other a third of a mile west of the west bluff of The Black Tusk. These were submitted for identification in 1947 to C. R. Stelck, then a Ph.D. candidate in paleontology at Stanford University. Stelck found sufficient material in the collection from the second

locality, notwithstanding the distorted and fragmentary preservation, to recognize two species, "... *Inoceramus* cf. *I. vancouverensis* Shumard and *Inoceramus* cf. *I. schmidti* Michael..." These can be compared with specimens collected from Sucia Island, 80 miles south of the occurrence in the Mount Garibaldi map-area. The former "species belongs to the widespread group of affiliated *Inocerami* referred to *I. balticus* and *I. barabini*, etc. occurring in the Senonian of Sucia Island, Japan, Europe, interior of North America". The latter resembles species found in the Senonian of Europe and Japan, notably a Japanese form *I. japonicus* var γ Nagao et Matumoto. "The Senonian age is well established."

Although correlative rocks of Sucia Island are part of the Nanaimo series (McLellan, 1927, p. 121–122, Buckham, 1947, p. 461–462, Usher, 1952, p. 57–17), extending for 125 miles northwestward, these two species of *Inoceramus* have not been reported at any other locality in the vicinity. Lithologic correlation of the Cheakmus Formation and some part of the Nanaimo series has not proven possible. No volcanic rocks are known in the Nanaimo series that might correspond to the Empetrum Formation.

No sedimentary successions of known Upper Cretaceous age have hitherto been reported from the southern Coast Mountains, and paleontologic correlations with rocks of other areas in this mountain chain are not possible.

The abundance of granitic debris in the Upper Cretaceous sediments of the map-area and its apparent absence in the older metavolcanic and metasedimentary rocks probably has local value as a basis for age determination. Similar granitic debris is abundant in known Lower Cretaceous rocks, and occurs locally in Upper or Middle Jurassic rocks of the southern Coast Mountains and adjacent areas (Duffel and McTaggart, 1952, p. 31); in still older rocks it is uncommon if present at all. The granite-bearing conglomerates of the Cheakamus Formation, formerly thought to be Triassic or older, have been used to support the contention that pre-Triassic plutonic rocks were once exposed in their vicinity (Schofield, 1941, and others), but such a conclusion must now be set aside. In the Britannia Beach map-area a few plutonic fragments have been reported in conglomerate of the Britannia Formation (LeRoy, 1908, p. 15) and in a lava flow of this same unit (James, 1929, p. 13), and boulders of feldspar and quartz-feldspar porphyries are present in a conglomerate in the lower Goat Mountain Formation (James, 1929, p. 16). These isolated occurrences are not sufficient in themselves to justify the assignment of a post-mid Mesozoic age to these two formations. The Mount Lomond arkose of this same area, however, is reported to contain an appreciable amount of detritus from granitic rocks (James, 1929, p. 9–10), and the possibility that this unit may correspond to the Cheakamus Formation should be investigated.

Plutonic Rocks—Part 2

Younger quartz diorites.—The younger quartz diorites occur as several separate masses in and adjacent to the Upper Cretaceous bedded rocks of the northern part of the map-area. The body which underlies at least 5 square miles near Empetrum Ridge may extend over a very much greater area and be genetically related to the gneisses northeast of Corrie Peak. The three other masses of younger quartz diorite large enough to be indicated on the map are exposed over a total area of only 0.45 square miles, but an undetermined proportion of the stock on the north side of Garibaldi Lake is concealed by water.

The younger quartz diorites are so similar lithologically to some of the varieties of the Cloudburst quartz diorites that, were it not for their relationships to the stratified rocks, no age distinction would have been recognized. All the younger quartz diorites are notably deficient in potassium feldspar, as are their older counterparts. Plagioclase is the predominant mineral in all specimens, and quartz is conspicuous. The chief original mafic mineral appears to have been hornblende, but in most places this has been altered to chlorite which only locally preserves the crystal outline of the former mineral. Feldspar grains, some of which are partly altered to epidote, generally present their original euhedral form. Quartz has not been notably brecciated. Discrete intersecting shear planes are common between the crystals of quartz and plagioclase and are rendered conspicuous by films of chlorite or sericite or both. Locally, where these shear planes are numerous and subparallel, the rock breaks readily into platy fragments or flakes; in other places where the shear planes may be widely spaced and intersect at large angles the rock breaks into irregular rhombohedral fragments. The stock on the north side of Garibaldi Lake is, however, undeformed, presents a simple joint pattern, and contains unaltered hornblende and biotite. The quartz diorite of sills

cutting the Helm formation around the margins of Helm Glacier and north of Helm Peak is also undeformed.

That the younger quartz diorites are clearly intrusive into the stratified Upper Cretaceous (?) rocks is shown by the tightly frozen contacts truncating the bedding of the wall rocks, by associated metamorphic zones, and in the case of the stock at Garibaldi Lake, by apophyses penetrating the adjacent strata. The sills in the Helm Formation have not been traced into the quartz diorite mass of the northern edge of the map-area and may or may not be offshoots of this intrusive.

The quartz diorite body at the south end of Empetrum Ridge has been intruded along a probable fault contact between the Empetrum and Cheakamus formations, and is itself somewhat sheared. The schistosity of this intrusive strikes parallel to its contacts and dips steeply northward roughly parallel to the schistosity in the bedded rocks on either side; it may reflect continued movement along the fault zone after emplacement of the quartz diorite.

The quartz diorite at the northern edge of the map-area intrudes the Empetrum and Helm formations, but its age relationship to the Cheakamus Formation is undetermined. It is faulted against this formation along part of the contact; the contact between quartz diorite and metavolcanic rocks farther northwestward may be an extension of the same fault. A broad tongue of quartz diorite extends along the line of inferred faulting between the Empetrum and Helm formations as if it intruded a preexisting fault zone. Secondary foliation in this quartz diorite mass, as much as 1½ miles from its contact, tends to parallel schistosity in the Helm and Empetrum formations. This body of quartz diorite, like the one south of Empetrum ridge, was probably intruded during an extended period of faulting and shearing.

The stock on the north side of Garibaldi Lake is, on the other hand, undeformed. It invades both the upper part of the Cheakamus Formation and the lower part of the Helm Formation with no sign of an offset along the inferred fault separating these two stratigraphic units.

Squamish batholith.—The Squamish batholith, a major pluton consisting essentially of fresh granodiorite and quartz monzonite, is exposed on both sides of lower Squamish Valley and upper Howe Sound. That part of the pluton within the map area is 10 miles long and up to 7 miles wide, and covers an area of almost 40 square miles. An outlier of the same rock, one-half by 1½ miles, is exposed in the eastern part of Squamish Valley 4 miles northeast of the town of Squamish. The "northwest area of batholithic rocks" in the Britannia Beach map-area (James, 1929, p. 53–54) is a part of the Squamish batholith.

As seen in hand specimen, the granodiorite and quartz monzonite contains abundant quartz, in grains up to 10 mm in diameter, conspicuously zoned plagioclase, white to pink untwinned feldspar, and biotite. The rock is homophanous and lacks gneissic banding. The texture is remarkably uniform from place to place, although near the margins of the plutons some finer-grained varieties and basic inclusions exist. Pegmatite dikes are present but rare.

Microscopic study reveals rounded and fractured grains of quartz making up 18–35 per cent of the rock, euhedral crystals of zoned oligoclase making up 30–55 per cent, either microcline or perthite or both as interstitial grains making up 5–28 per cent, and biotite, with or without minor hornblende, making up 5–20 per cent. Accessory minerals include apatite, sphene, and black opaque material; muscovite and epidote occur in very minor amounts as alteration products.

Little is known of internal structure. No primary foliation has been recognized, and such variations in composition as have been noted cannot be related to distance from the walls of the intrusive rock or to altitude within it. Jointing is simple, consisting of three or four joint sets intersecting at large angles to form conspicuous rectangular or triangular patterns on exposed surfaces. Joints are as a rule widely spaced, in places tens of feet apart. The pattern has apparently been developed by a simple arrangement of stresses in a single period of fracturing, probably during the initial cooling and with little or no externally applied force. The simple jointing and the lack of any detectable interruption in the smooth northeastern contact of the batholith suggests no faulting after intrusion.

The Squamish batholith has not significantly altered or deformed its wall rocks. Cloudburst quartz diorite at its contact with the Squamish batholith between Squamish River and Fries Creek is typical but for a lack of hornblende; instead greenish biotite and some chlorite in large aggregates of fine-grained flakes make up about 20 per cent of the rock. The bedded rocks of Goat Ridge within a few hundred feet of the batholith are flat lying and apparently undisturbed by faulting. Some partly assimilated masses of these older rocks are found within the marginal part of the batholith.

A characteristic topographic expression is imparted to the granodiorite and quartz monzonite by the scarcity of jointing and the simplicity of the joint pattern. Long smooth cliff faces with hardly a foothold for vegetation are common, and some of the valley walls rival in steepness and height those of Yosemite Valley in California. The narrowness and depth of the fiord valley of Howe Sound may be attributed for the most part to the joint control of glacial plucking, although the smoothness of some exposures has been enhanced by glacial abrasion. Postglacial streams and weathering have had little effect on the rock and, although it is susceptible to disintegration, many ice-polished surfaces remain. Talus piles are small even though characterized by blocks of enormous size.

Direct evidence of the age of the Squamish batholith is lacking, except that it postdates the highly fractured Cloudburst quartz diorites and the Goat Mountain Formation of James (1929). The absence of marked deformation during and after intrusion, however, strongly suggests that it is younger than the tilted and block-faulted Upper Cretaceous sedimentary rocks 15 miles to the north.

Castle Towers batholith.—The Castle Towers batholith, made up of quartz diorite and minor trondhjemite, is situated near the northeastern corner of the map-area. About 20.7 square miles of the batholith lies within the map-area, but 7.7 square miles is concealed by glaciers. Its extent beyond the boundaries of the map-area are imperfectly known, but recent reconnaissance mapping by the Geological Survey of Canada (Roddick and Armstrong, 1957) plus indications of structure in distant views suggests it extends without interruption eastward 8 miles beyond the map-area and southward to the northern slope of Mamquam Mountain. If so, it occupies a roughly circular area about 10 miles in diameter and about 85 square miles in area.

Most of the mapped area of the batholith consists of quartz diorite, massive and homophanous except near the margins where a faint flow structure, shown either by preferred orientation of minerals or by successive lighter and darker bands, parallels the steeply dipping walls. One sample, from the ridge north of Sphinx Mountain and about half a mile from the western contact, consists of about 60 per cent oligoclase, 20 per cent quartz, 10 per cent hornblende, and smaller amounts of orthoclase, biotite, sphene, epidote, apatite, sericite, and black opaque grains. Other specimens of quartz diorite differ from this one macroscopically chiefly in the proportions of light and dark minerals; the rocks tend to be richer in mafic minerals near the margins of the batholith than toward the center.

A white rock consisting essentially of oligoclase and quartz with minor orthoclase occurs on Parapet Peak. This rock, unlike anything hitherto described from the Coast Mountains except perhaps the "sodic granodiorite" of the Vancouver area (Phemister, 1945, p. 67), approaches in chemical and mineralogical composition Goldschmidt's trondhjemite. Like the original trondhjemite, this quartz-rich rock is notably deficient in potassium feldspar, although the amount present in the mode, approximately 10 per cent, is more than in Goldschmidt's specimens. Roughly rectangular oscillatory-zoned crystals of oligoclase, many with corroded or myrmekitic borders, make up about 60 per cent of the rock, and flakes of biotite less than 5 per cent. Interstitial grains, about 35 per cent of the rock, consist predominantly of quartz and minor orthoclase. Accessory minerals include sphene, apatite, black opaque grains, and clinozoisite; alteration products present in very minor amounts include sericite and chlorite.

A few well-defined dikes cut both quartz diorite and trondhjemite and consist of fine-grained basic and acid rocks as well as pegmatites. A specimen of pegmatite from Parapet Peak consists of a graphic intergrowth of quartz and microcline with minor muscovite and garnet.

The contact between trondhjemite and quartz diorite is concealed by glacial ice at the one point where it was crossed, but the gradual decrease in the dark mineral contact of the quartz diorite as the contact is approached suggests that the two rocks may grade into one another. Recent mapping (Roddick and Armstrong, 1957) indicates that the more acid rock underlies an elliptical area 3 miles wide extending 7 miles southeast from Parapet Peak, and thus occupies about 16 square miles or almost 20 per cent of the area of the pluton.

Regular jointing is characteristic. Three joint sets are typical and define roughly rectangular blocks. Two of the joint sets are steeply dipping, one nearly parallel to the nearest part of the batholith wall, the other nearly normal to it. The strike of the third set is almost parallel to the margin of the intrusive rock and the dip is about 15° toward this margin. These attitudes persist in the entire northwestern part of the batholith with

but slight variations and conform with a radial symmetry. No displacement has been observed on the steeply dipping joints, but thin pegmatite dikes occupying a few of the gently dipping joints in quartz diorite show slickensides paralleling the direction of dip. None of these fault planes has been seen cutting banded quartz diorite, hence the direction and amount of movement is undetermined.

The Castle Towers bath lith is bounded on the northwest, west, and southwest by a belt of metamorphic rocks several hundred feet to almost half a mile wide. North of Sphinx Mountain these metamorphic rocks consist mainly of medium- to fine-grained biotite gneiss which grades with no apparent break into the Cloudburst quartz diorites. Microscopic study of the gneiss reveals numerous small undeformed biotite flakes, showing a strong preferred orientation, in a matrix of oligoclase, having randomly oriented twin lamellae, and quartz, also lacking obvious preferred orientation. If this rock is a metamorphic derivative of the Cloudburst quartz diorites, it was first subject to deformation, by which the original coarse texture was reducted and the orientation pattern of the biotite developed, and was then recrystallized, enabling quartz, feldspar, and biotite to reform in an unstrained state. Unoriented porphyroblasts of hornblende in the gneiss west of Sphinx Mountain provide additional evidence of post-tectonic recrystalliztion. South of Sphinx Mountain the Cloudburst quartz diorites, only locally deformed, are separated from the Castle Towers batholith by a belt of amphibolite, biotite schist, and migmatite, much of it apparently derived from volcanic and sedimentary rocks. Foliation in the immediate vicinity of the Castle Towers batholith approximately parallels its walls. Vertical lineation is locally developed. No diagnostic minerals, other than biotite, have been found to indicate the grade of metamorphism.

The boundary between these metamorphic rocks and the Castle Towers batholith is sharp, and xenoliths of the former or apophyses of the latter are lacking. The contact, both in small exposures and on valley walls, is nearly vertical. At one place west of Castle Towers Mountain the contact is displaced several feet by an eastward-dipping reverse fault, a marginal thrust associated with the intrusion of the batholith. No other faults, large or small, have been observed at the contact and, although scattered upthrusts of the scale of the one observed could readily be missed, no major faults with displacements comparable to those in the near-by Upper Cretaceous rocks are likely to be undetected.

Notable divergences between foliation in the metamorphic rocks and the adjacent contact are known. In one locality 2½ miles northeast of Castle Towers Mountain the trend of banding in amphibolite veers sharply northeastward, northward, then northwestward away from the contact to describe a vertical fold; one limb parallels the contact, the other the regional trend. Similar variations in attitude have been seen south of Viking Ridge where steeply dipping foliation strikes southeastward, southward, and southwestward with increasing distance from the batholith. Between these two localities the regional trend of structure is nearly parallel to the local trend of the contact, and variations in the attitude of foliation are neither observed nor expected.

The batholith has apparently escaped all external deformation since its intrusion judging from the simple joint patterns and the lack of major faults.

Direct evidence of the age of the plutonic rock is limited. It cuts Cloudburst quartz diorites, of pre-Upper Cretaceous age. Most significant is the lack of external deformation, suggesting intrusion after the period of block faulting which affected the Upper Cretaceous sedimentary rocks only a few miles away.

Chemical compositions.—The compositions of the various batholithic rocks, early and late, are indicated by the chemical analyses of six representative rocks from the map-area (Table 3). These six analyses and four analyses from the Vancouver area (Phemister, 1945, p. 66) are illustrated graphically on a Harker variation diagram (Fig. 2). Although rocks of at least two ages are included, nearly all the points fall on a single set of smooth curves. Only the Castle Towers trondhjemite (SiO_2 = 71.2) and the Caulfeild "sodic granodiorite" (SiO_2 = 70.0) depart significantly from the curves; both are distinctly richer in Na_2O and CaO and poorer in K_2O and Fe_2O_3 than the average for corresponding silica contents. With one exception the rocks from the Vancouver area are slightly richer in K_2O than the corresponding rocks of the Garibaldi area.

The analyses also demonstrate the marked range in composition possible within a single pluton, as for example the variation from 54.6 per cent SiO_2 to 69.4 per cent SiO_2 within the Cloudburst quartz diorite.

The alkali-lime index, that is the silica content at which CaO = K_2O + Na_2O (Pea-

TABLE 3.—CHEMICAL ANALYSES AND NORMS OF THE PLUTONIC ROCKS
All analyses by W. H. Herdsman, Glasgow, Scotland

	1	2	3	4	5	6
SiO_2	65.29	54.60	69.37	71.23	60.42	73.63
Al_2O_3	16.86	18.97	16.48	16.09	17.39	14.12
FeO	2.24	4.96	1.83	1.23	3.55	1.75
Fe_2O_3	1.96	2.34	.94	.47	1.90	.92
TiO_2	.46	.84	.32	.21	1.02	.24
MnO	.14	.13	.12	Traces	.12	Traces
CaO	5.18	8.08	3.23	3.88	7.26	1.62
MgO	2.06	4.35	1.02	.50	2.81	.51
K_2O	1.12	.42	1.36	1.57	1.03	3.04
Na_2O	3.18	3.39	3.37	4.44	3.62	3.52
H_2O, −105°C	.12	.07	.07	.05	.16	.07
H_2O, +105°C	1.43	1.45	1.92	.32	.46	.43
CO_2	Nil	Nil	Nil	Nil	Nil	Nil
P_2O_5	.17	.33	.13	.08	.22	.06
Total	100.21	99.93	100.16	100.07	99.96	99.91
Specific Gravity	2.74	2.82	2.67	2.65	2.78	2.63
Quartz	27.93	8.25	35.47	29.90	16.12	36.55
Orthoclase	6.62	2.48	8.04	9.27	6.08	17.96
Albite	26.89	28.67	28.49	37.55	30.61	29.76
Anorthite	24.58	35.31	15.18	18.72	28.17	7.64
Corundum	1.41	.....	3.90	.23	.....	2.24
Diopside	.....	2.13	.....	.....	5.19	.....
Hypersthene	7.13	15.83	4.82	2.77	7.97	3.33
Magnetite	2.84	3.39	1.36	.68	2.75	2.24
Ilmenite	.87	1.59	.60	.40	1.94	.46
Apatite	.39	.76	.30	.18	.51	.14
H_2O+	1.43	1.45	1.92	.32	.46	.43
H_2O-	.12	.07	.07	.05	.16	.06
Total	100.21	99.93	100.16	100.07	99.96	99.91
MgO/FeO in mafics	72/28	64/36	53/47	45/55	67/33	38/62
Ab/An in feldspar	52/48	45/55	65/35	67/33	52/48	80/20

1. Cloudburst quartz diorite, 1 mile north of Cheakamus station.
2. Cloudburst diorite, south end of Cheakamus Bridge, P.G.E. Ry.
3. Cloudburst quartz diorite, northwest side of Lake Lucille.
4. Castle Towers batholith—trondhjemite, Parapet Peak.
5. Castle Towers batholith—quartz diorite, half a mile northeast of Sphinx Mountain.
6. Squamish batholith—granodiorite, three-quarters of a mile northeast of Squamish.

cock, 1931), is 65.5 in the Garibaldi rocks, one of the highest values on record and in accord with the generally high lime content of other igneous rocks in the western part of the North American cordillera. For comparison the alkali-lime index of the Katmai suite is 63.8 (Peacock, 1931), of the Mount Lassen suite 63.9 (Williams, 1932), of the Oregon intrusive rocks 61.6 (Buddington and Callaghan, 1936), and of the Sierra Nevada intrusive rocks 60.8

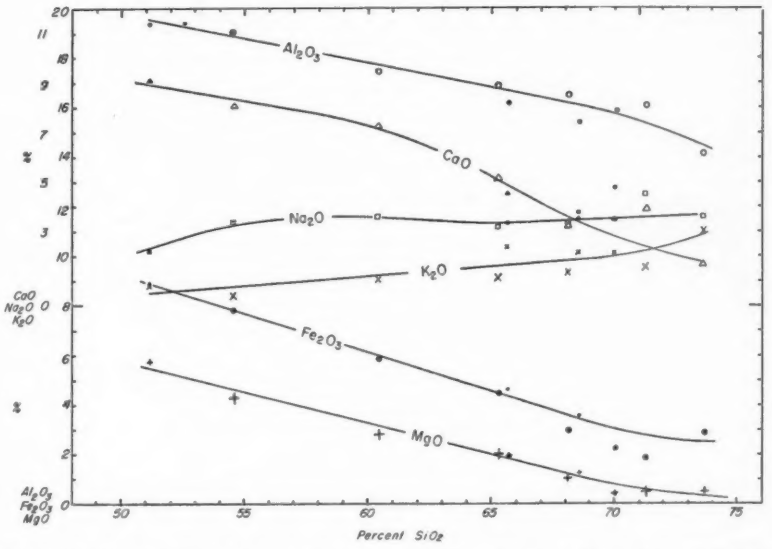

FIGURE 2.—VARIATION DIAGRAM FOR THE COAST INTRUSIVE ROCKS OF THE MOUNT GARIBALDI MAP-AREA (LARGE SYMBOLS) AND THE VANCOUVER AREA (SMALL SYMBOLS)

(Buddington and Callaghan, 1936). The alkali-lime index of the Cenozoic lavas of the North Island of New Zealand is 62.7 (Benson, 1941), and MacGregor's data (1938) on the composition of West Indian lavas when plotted suggest an alkali-lime index as high as 66.5.

Specific gravities, as listed in Table 3, provide an approximate index to the chemical compositions of rocks in the Garibaldi suite. If the data are plotted on a silica content-density diagram a straight line can be drawn such that no point departs from it by more than 2 per cent in silica content or more than 0.02 in specific gravity.

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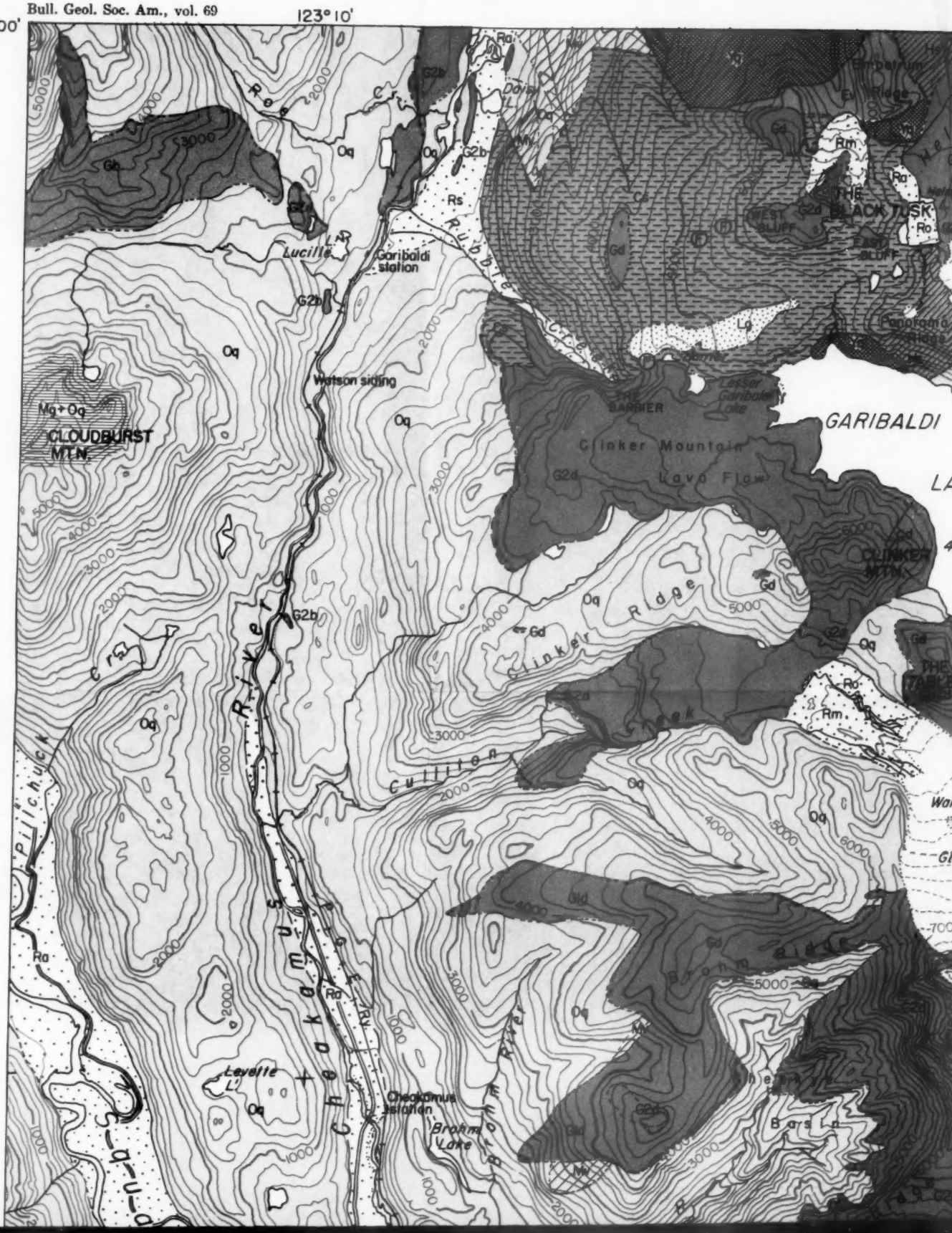
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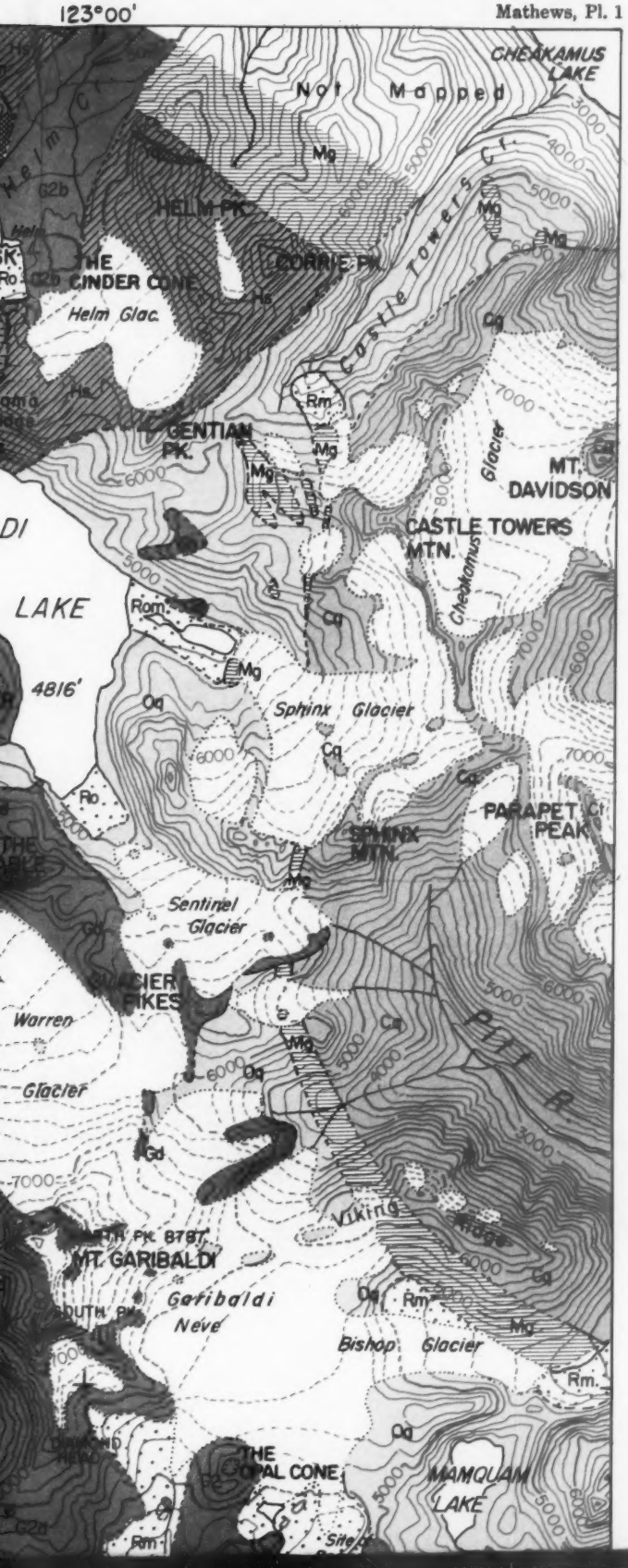
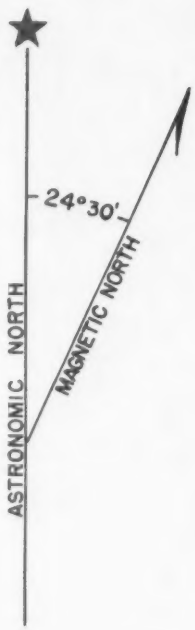
THE BLACK TUSK
WEST BLUFF
EAST BLUFF
Rainbow Ridge
Panorama Ridge
Lesser Garibaldi Lake
GARIBALDI
THE BARRIER
Clinker Mountain Lava Flow
CLINKER MTN.
Clinker Ridge
Clinker Creek
Cutline
Pilchuck
Squamish River
Squamish
Lucille
Garibaldi station
Watson siding
CLOUDBURST MTN.
Levette L.
Cheakamus station
Brohm Lake
Brohm River
Cheakamus River
Bars La
Pubble Creek
Rs
Oq
Mq+Oq
G2b
G2d
Gd
Ra
Rm
Rv
Ro
Ls
Qg





LEGEND

QUATERNARY

Glacier — (1947 limits)

Ra, Rm — Recent alluvium – a, slide debris – s, outwash – o, moraine – m

Lg, Lc — Late Glacial terrace gravels – g, clays – c, slide debris – s

GARIBALDI GROUP

Late Glacial and early postglacial – 2, older – 1; rhyodacite – r, dacite – d, andesite and basalt – b

PROBABLY POST-UPPER CRETACEOUS YOUNGER INTRUSIONS

Cq, Cr — CASTLE TOWERS BATHOLITH
Quartz diorite – q, trondhjemite – t

Sg — SQUAMISH BATHOLITH
Granodiorite, quartz monzonite

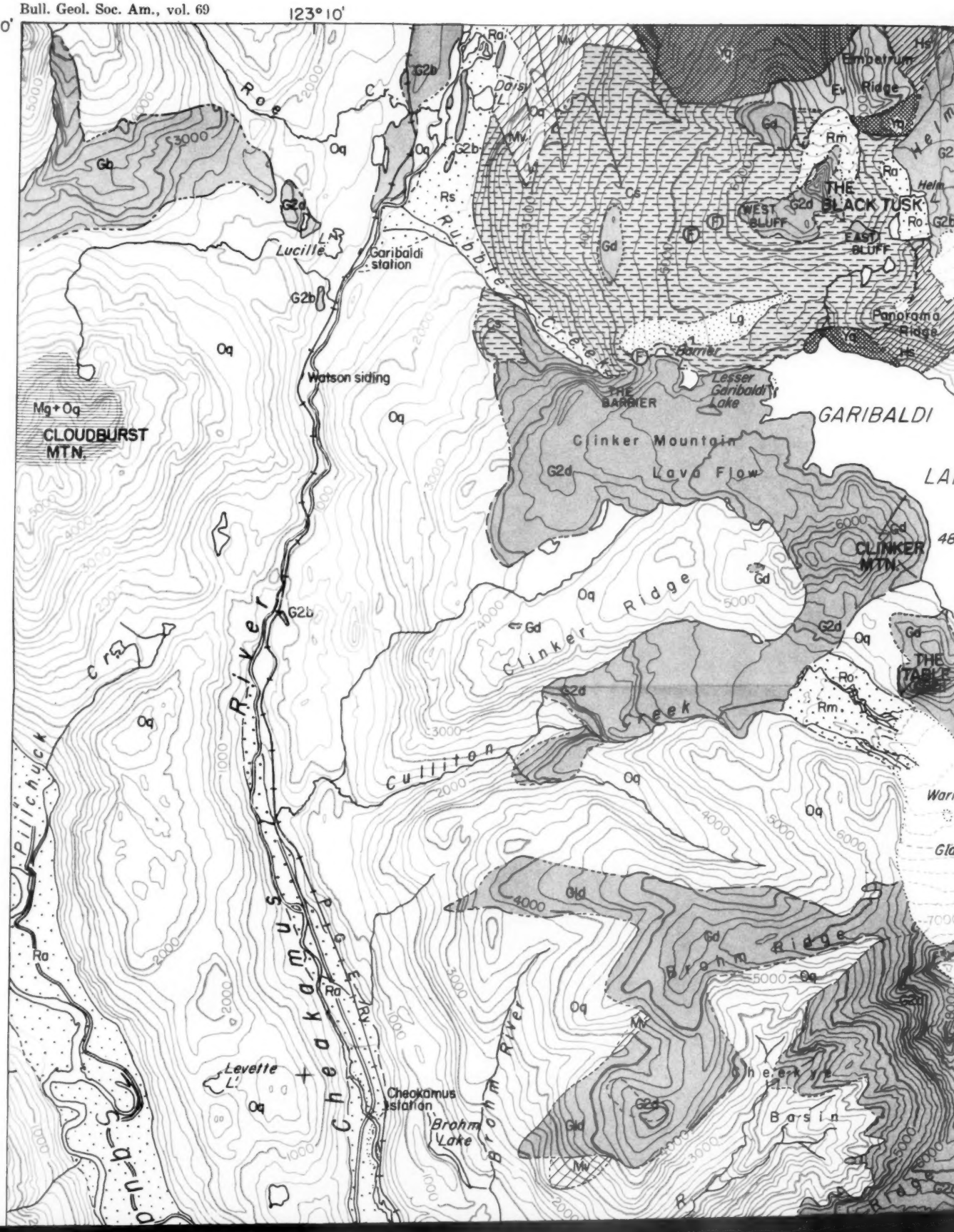
Yq — YOUNGER QUARTZ DIORITES

UPPER CRETACEOUS

HELM FORMATION
Slaty argillite, quartzite, minor conglomerate and calcareous beds

EMPETRUM FORMATION
Metavolcanic rocks, conglomerate, limestone

CHEAKAMUS FORMATION


123°10'
50°00'
49°50'
Roe Cr
Lucille L.
Garibaldi station
Rubble Creek
Watson siding
CLOUDBURST MTN.
Mg + Oq
Pilchuck Cr
Pilchuck River
Levette L.
Cheakamus River
Cheakamus station
Brohm Lake
Brohm River
Culliton Creek
Clinker Ridge
Clinker Mountain Lava Flow
CLINKER MTN.
THE BARRIER
Lesser Garibaldi Lake
GARIBALDI
Panorama Ridge
Empetrum Ridge
THE BLACK TUSK
WEST BLUFF
EAST BLUFF
Daisy L.
Brohm Ridge
Cheakamus Basin
Gb
Oq
Rs
Gd
G2d
G2b
Ra
Rm
Mv
Cs
Lg
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Hs
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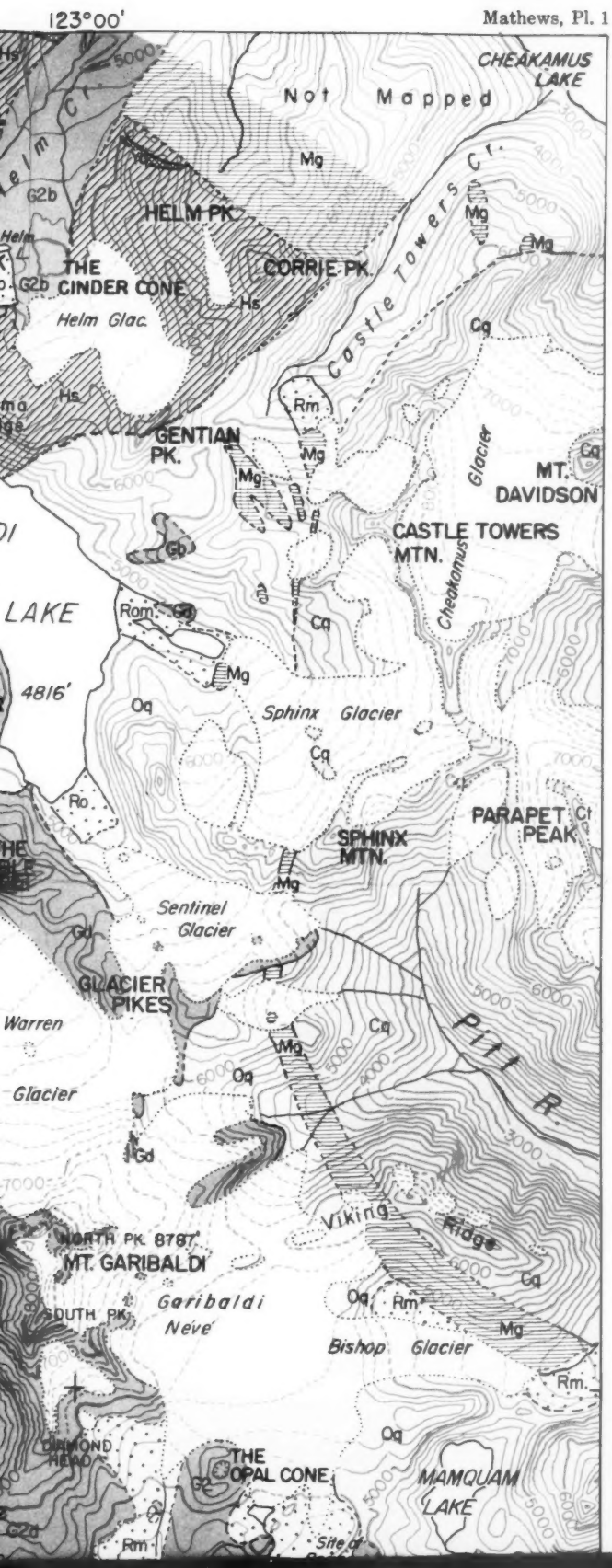

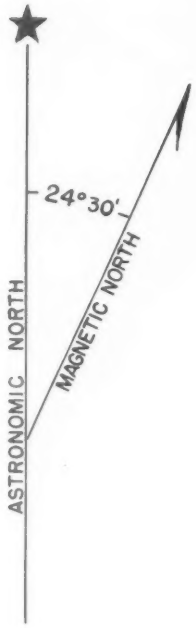

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Lg Lc — Late Glacial terrace gravels – g, clays – c, slide debris – s

GARIBALDI GROUP

G2d G1b — Late Glacial and early postglacial – 2, older – 1; rhyodacite – r, dacite – d, andesite and basalt – b

PROBABLY POST-UPPER CRETACEOUS

YOUNGER INTRUSIONS

Cq Ct — CASTLE TOWERS BATHOLITH
Quartz diorite – q, trondhjemite – t

Sg — SQUAMISH BATHOLITH
Granodiorite, quartz monzonite

Yq — YOUNGER QUARTZ DIORITES

UPPER CRETACEOUS

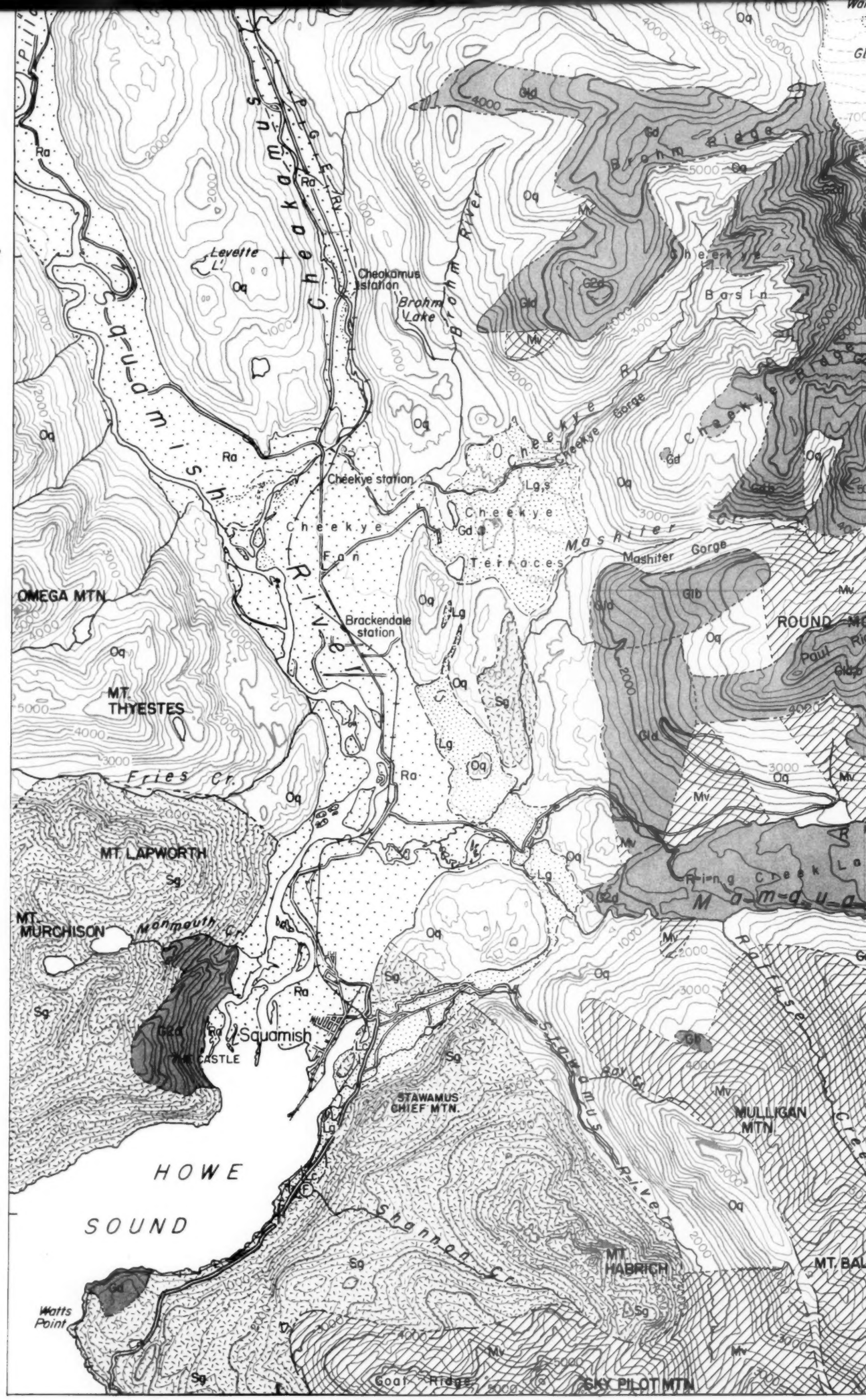
HELM FORMATION

Hs — Slaty argillite, quartzite, minor conglomerate and calcareous beds

EMPETRUM FORMATION

Ev — Metavolcanic rocks, conglomerate, limestone

CHEAKAMUS FORMATION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Squamish
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Cheekye station
Brackendale station
Cheekye Fan
Cheekye Terraces
Cheekye Gorge
Mashiter Gorge
Brohm Ridge
Brohm Lake
Cheakamus station
Levette L.
Squamish River
Cheakamus River
Pacific Rv.
Ring Creek
Rainbow Creek
Stawamus River
Stawamus Bay Cr.
Shannon Cr.
Fries Cr.
Monmouth Cr.
Goat Ridge
OMEGA MTN.
MT. THYESTES
MT. LAPWORTH
MT. MURCHISON
STAWAMUS CHIEF MTN.
MT. HABRICH
MULLIGAN MTN.
SKY PILOT MTN.
ROUND MTN.
HOWE SOUND
Watts Point
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Ra
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Glc
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Gd

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UPPER CRETACEOUS

Hs — HELM FORMATION
Slaty argillite, quartzite, minor conglomerate and calcareous beds

Ev — EMPETRUM FORMATION
Metavolcanic rocks, conglomerate, limestone

Cs — CHEAKAMUS FORMATION
Graywacke, conglomerate, arkose, minor argillaceous and calcareous beds

PRE-UPPER CRETACEOUS

OLDER INTRUSIONS

Oq — CLOUDBURST QUARTZ DIORITES
Foliated and unfoliated quartz diorite, minor diorite etc.

Mv — METAVOLCANIC AND METASEDIMENTARY ROCKS
Greenstone, slaty argillite, breccia, minor conglomerate, limestone, green schist

UPPER AND PRE-UPPER CRETACEOUS

Mg — Migmatite, amphibolite, biotite schist, paragneiss

Contact — location definite
Contact — location uncertain
Fault – located
Fault – inferred
Ⓕ Fossil locality
Railway
Public road

Topography from B.C. Dept. of Lands and Forests manuscript maps. Geology by W.H. Mathews — 1946 – 47. Revised 1957

INDEX MAP (Scale: 500 miles)

TISH COLUMBIA

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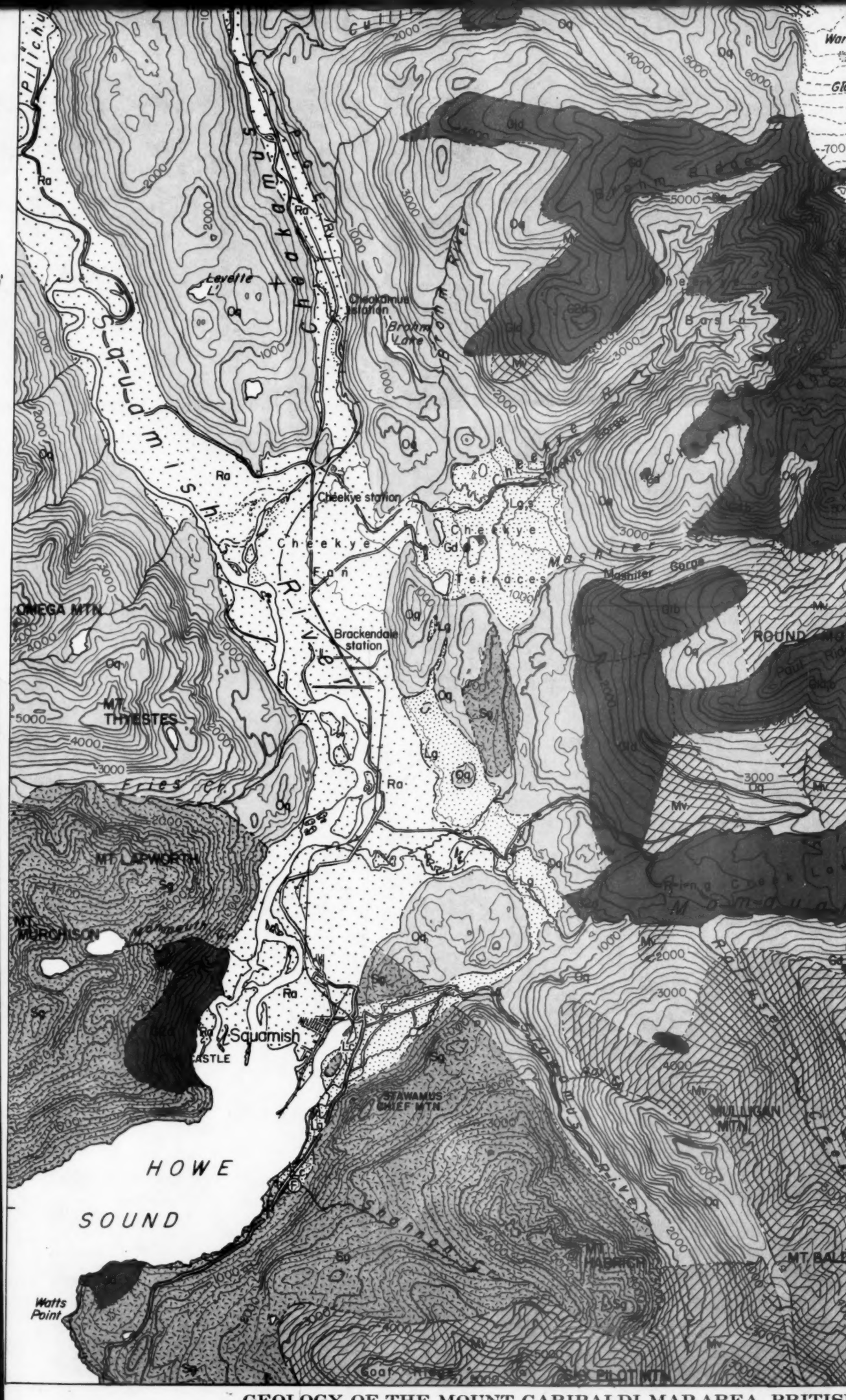
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MT. LAPWORTH
MT. MURCHISON
Squamish
STAWAMUS CHIEF MTN.
HOWE SOUND
Watts Point
MULLIGAN MTN.
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Levette L.
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TISH COLUMBIA

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GARIBALDI GROUP

Late Glacial and early postglacial – 2, older – 1; rhyodacite – r, dacite – d, andesite and basalt – b

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Slaty argillite, quartzite, minor conglomerate and calcareous beds

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Metavolcanic rocks, conglomerate, limestone

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Graywacke, conglomerate, arkose, minor argillaceous and calcareous beds

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Public road

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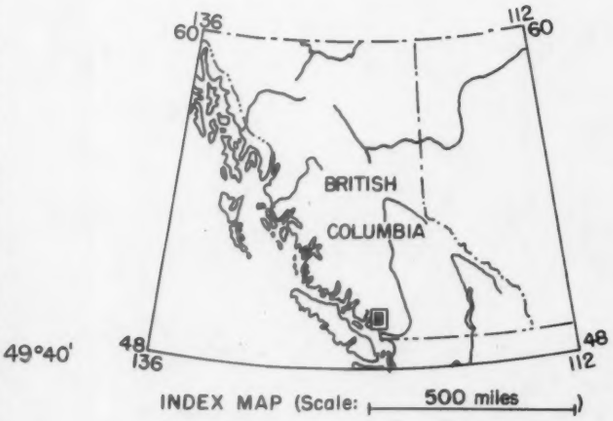

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 69, PP. 179–198, 5 FIGS., 6 PLS. FEBRUARY 1958

GEOLOGY OF THE MOUNT GARIBALDI MAP-AREA, SOUTHWESTERN BRITISH COLUMBIA, CANADA

PART II: GEOMORPHOLOGY AND QUATERNARY VOLCANIC ROCKS

BY W. H. MATHEWS

ABSTRACT

The Mount Garibaldi map-area is in the southern Coast Mountains of British Columbia. The bold topography of the area is a product of two cycles of stream erosion modified by continental and alpine glaciation and by Quaternary vulcanism.

The volcanic rocks of the map-area are lavas and pyroclastic rocks which postdate the second cycle of stream erosion; most if not all are Pleistocene, and many appear to have been erupted during periods of glaciation. Many of the anomalous structures can be attributed to deposition of lava or pyroclastic debris against, beneath, or onto glacial ice. Extrusive rocks range in composition from basalt to dacite; the latter are most voluminous.

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INTRODUCTION[1]

Previous Work

Reference has been made in the literature to the volcanic rocks and geomorphology of the map-area by R. W. Brock (1928), H. T. James (1929), and F. C. Bell *et al.* (1932). E. M. J. Burwash included reconnaissance observations from the Garibaldi Lake area in his doctorate dissertation (1918) as well as in earlier papers (1914a; 1914b). Additional brief notes on the geology of various parts of the

[1] Location and access and field work are described in Part I: Igneous and Metamorphic Rocks.

map-area are scattered through the publications of the British Columbia Mountaineering Club (1908–) notably in its book *The Northern Cordilleran* (1913), and in the Canadian Alpine Journal (1906–)

Acknowledgments

In addition to those who assisted in this study mentioned in Part I: Igneous and Metamorphic Rocks, Mr. L. R. Harrison, a pioneer in the studies of Quaternary volcanic rocks northwest of Cheakamus River, provided valuable information on their distribution and petrography.

Geomorphology

General Statement

Topography is bold throughout the map-area, and the relief is great. The highest peak is Mount Garibaldi, 8787 feet above sea level, and the lowest elevation is in Howe Sound, the floor of which lies about 800 feet below sea level. In spite of this great relief ridges and summits in most parts of the map area are rounded, and only at altitudes of more than 5500 feet above sea level in the south and more than 7000 feet in the north are sharp peaks and arêtes common. Precipitous cliffs are generally confined to northward-facing slopes and to higher levels.

Three major elements to the topography can be recognized: (1) an erosion surface of moderate relief, truncating the metamorphic and plutonic rocks and the Upper Cretaceous bedded rocks, now represented by a few plateaulike ridge tops and here referred to as the upland surface, (2) a younger erosion surface represented by the relatively steep walls and hummocky floors of the deeper valleys and here referred to as the valley surface, and (3) constructional surfaces developed by volcanic, fluvial, and glacial processes. A few areas exist which might be assigned to either the upland or the valley surface, and many of the constructional surfaces have become more or less modified by erosion, hence it is not possible to recognize clear-cut boundaries between these three elements in all parts of the map area.

Upland Surface

Identifiable relics of the upland surface bevelling the older rocks are rare, of limited size, and confined to the central and northern part of the map-area. The best examples are Clinker Ridge, where a triangular area of between 3 and 4 square miles of subdued topography still remains between the 4000- and 6000-foot contours, and the broad summit of Brew Mountain, 2 miles north of the map-area, where more than a square mile of rolling upland exists between the 4500- and 5700-foot levels. Although bluffs are common in these areas, the land surface rarely falls more than 1000 feet in a mile. Similar, though smaller, upland relics occur on the north slope of Round Mountain, between Clinker Mountain and The Table, and northeast of Empetrum Ridge. In several places, as for example near the crest of Round Mountain, the upland surface is party buried by later volcanic rocks.

This old subdued land surface, which must once have extended over a large part, if not all, of the map area, was apparently produced by long-continued erosion at a time when the area stood from 3000 to 4000 feet lower than it does today. How low a relief was developed during this period of erosion cannot be determined surely, for some differential warping may since have taken place. Some slopes, such as that of Clinker Ridge which now declines 2000 feet in 3 miles, may have originated in part through the gentle tilting of an originally nearly horizontal surface.

No upland remnants exist today within the map-area south of Mamquam River, but with the exception of Sky Pilot Mountain and its immediate neighbors no peak or ridge rises above the 5600-foot level. Here the valley cutting has advanced to such a stage that all interstream areas have been reduced to sharp-crested ridges and the upland surface, assuming it once extended over this portion of the map-area, must have lain somewhat above the present summit level.

Peaks several miles both to the east and west of the known upland remnants have altitudes of more than 8000 feet and rise 2000 feet and more higher than the uplands previously mentioned. Conceivably these mountainous areas stood in the past, as they do today, high above the upland erosion surface, but more likely they represent areas where the upland has been uplifted to greater heights. Three conspicuous massifs stand amid the higher mountains in and to the east of the map area: (1) the Castle Towers massif (Pl. 2, Fig. 2), in the northeastern corner of the map area an area 1½ by 2 miles entirely above the 6000-foot contour, (2) the Mamquam massif (Fig.

3), immediately east of the map area, an area 2 by 3 miles entirely above the 6500-foot contour, and (3) an unnamed massif 10 miles east of Mamquam Mountain which embraces an area 5 by 9 miles most of which is above the 6000-foot contour. Had these three massifs existed during the upland erosion cycle as broad monadnocks rising 3000–5000 feet above the broad valley floors, they should have been dissected by at least one or two deep valleys. Even the present cycle, although much less advanced than its predecessor, has left very few interstream remnants of these dimensions. The lack of any deep valley in these massifs offers strong evidence against the monadnock hypothesis. According to the hypothesis favored here, the upland surface has been differentially uplifted, in these places as much as 6500 feet, to a level at or above the summits of these massifs. Because of their great height these massifs now support active glaciers which have so sculptured the landscape that no part of the upland surface itself remains.

Valley Surface

The term valley surface is applied to all the relatively youthful erosion forms which have developed during the current cycle: the walls of all the major valleys and the gently sloping land of valley floors and of benches. Hillsides representing this surface generally slope 2000–3000 feet in a mile and thus provide a marked contrast to the much lower slopes of the upland surface. On areas of soft or highly fractured rocks, however, the valley surface may have a gradient as little as 1200 feet in a mile, and the higher parts of such a slope can be hard to distinguish from the adjoining upland. Slopes falling more than 3000 feet in a mile are not rare, and in the valley of Pitt River and on the west slope of Squamish Valley slopes dropping more than 4000 feet in this same distance can be found. Steeper slopes of shorter length are common, and the most spectacular cliff, the northwest face of Stawamus Chief Mountain, drops 1200 feet at an average slope of more than 70°.

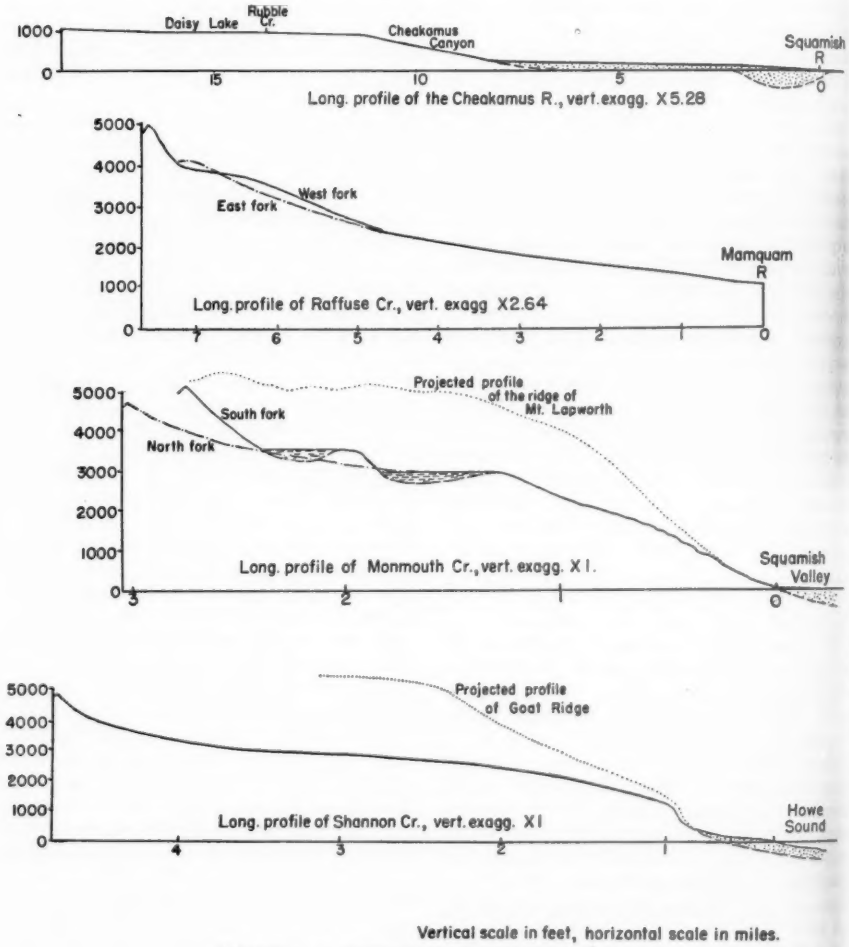
All parts of the valley surface have been somewhat modified by glacial erosion. V-shaped valleys with smoothly graded long profiles are rare: Raffuse Creek (Fig. 1; Pl. 1, fig. 2) provides the best example. Parts of the valleys of Mashiter and Culliton Creeks and of Cheekye, Brohm, and Mamquam Rivers do show stream-cut characteristics. Valleys heading in the high mountains, which were the source of much Pleistocene ice, and those paralleling the direction of ice dispersal, however, have been distinctly modified (Pl. 1, fig. 1; Pl. 2, fig. 1).

Longitudinal profiles of many valleys show marked overdeepening in their upper parts. Notable in this respect is the valley of Pitt River which within the first 3 miles of its 50-mile course to the edge of the Coast Mountains has been cut down to an altitude of only 2800 feet, more than 4000 feet below the adjacent peaks.

Irregularities of the longitudinal profiles are common. For example Cheakamus and Monmouth Valleys (Fig. 1) show stepped profiles which would be even more striking were the bedrock floors fully exposed.

Hanging valleys are well developed in the southwestern part of the map-area and in the Tantalus Range to the west of it. Valley floors commonly lie from 3000 to 5000 feet above sea level and from 2000 to 3000 feet below the near-by peaks. The mouths of the hanging valleys of Monmouth and Shannon Creeks enter the trunk valley of Squamish River-Howe Sound 1000 feet above sea level and possibly twice that height above the rock floor of the main valley. The profiles show no accordance with present sea level or with any other common level. Conceivably the hanging valleys were developed during the upland erosion cycle and survived later erosion because of resistance of their rock floors, especially where these consist of sparsely jointed Squamish granodiorite. This hypothesis, however, does not hold for valleys cut in the less resistant Cloudburst quartz diorites which were sufficiently weak to permit development of subdued and presumably old-age topography elsewhere in the map-area during the upland cycle. Moreover, since depth of these valleys and steepness of their walls are comparable in both Squamish granodiorite and Cloudburst quartz diorites, they are more probably later than the upland surface and formed in part by glacial action.

Cross sections of most valleys show a departure from the ideal form of stream-cut valleys, and only Raffuse Creek valley and Cheekye and Mashiter Gorges display well-developed V-shaped cross profiles (Pl. 1, fig. 2). Others, notably Pitt Valley, have U-shaped cross sections at least near the valley bottoms. The floors of most of the deeper valleys have been concealed by talus, alluvium, or lava, but Cheakamus Valley for 5 miles south of



Vertical scale in feet, horizontal scale in miles.

FIGURE 1.—LONGITUDINAL PROFILES OF STREAMS
Vertical scale in feet, horizontal scale in miles.

Daisy Lake is rock-floored and presents a hummocky surface with many knolls, like enormous *roches moutonnées*, rising as much as 500 feet above their surroundings. A similar topography exists on the east side of Squamish Valley near Mamquam River (Pl. 1, fig. 1).

The slope of the valley walls can be related to the extent of glacial modifications of valley

PLATE 1.—VALLEY FORMS

FIGURE 1.—Howe Sound from the northeast. Alluvial flats of Squamish River at right, hummocky rock floor of Squamish Valley in center, precipitous cliff of Stawamus Chief Mountain at left.

FIGURE 2.—Raffuse Creek Valley from the north illustrating the V-shaped cross profile.

PLATE 2.—GLACIAL FORMS

FIGURE 1.—Glacially modified valley of Castle Towers Creek looking northeast. Moraine of Castle Towers Glacier in foreground.

FIGURE 2.—Castle Towers Mountain from the north. Castle Towers Glacier in center of the photo with the 19th century moraine and the 1946 ice limit (dotted line). Mount Garibaldi in right rear and Mamquam Mountain in left rear. Royal Canadian Air Force photograph.

FIGURE 1

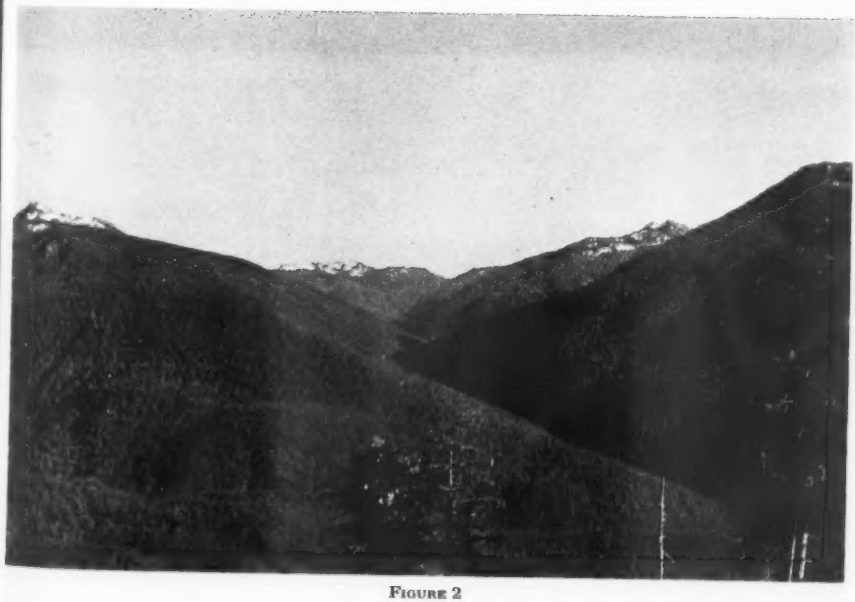


FIGURE 2

VALLEY FORMS

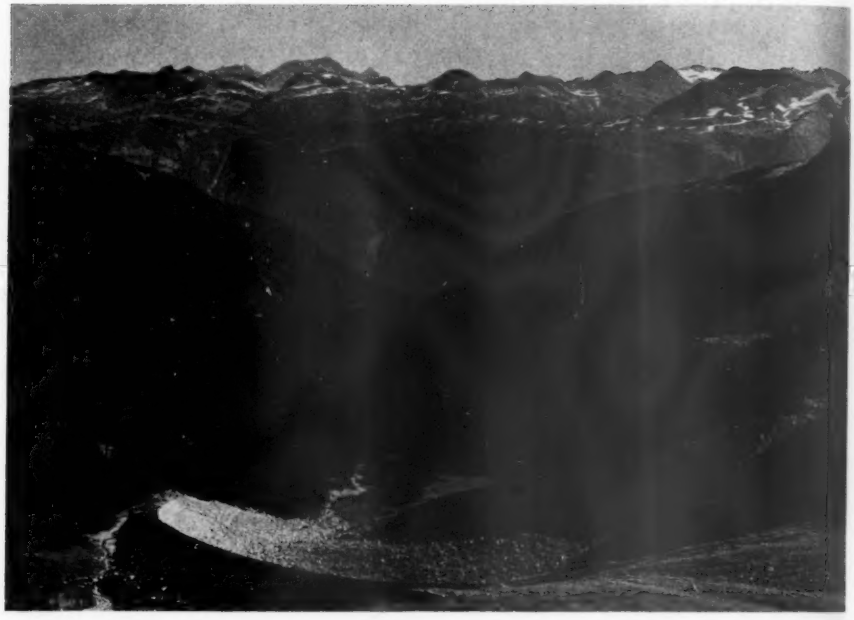


FIGURE 1

FIGURE 2

GLACIAL FORMS

FIGURE 1

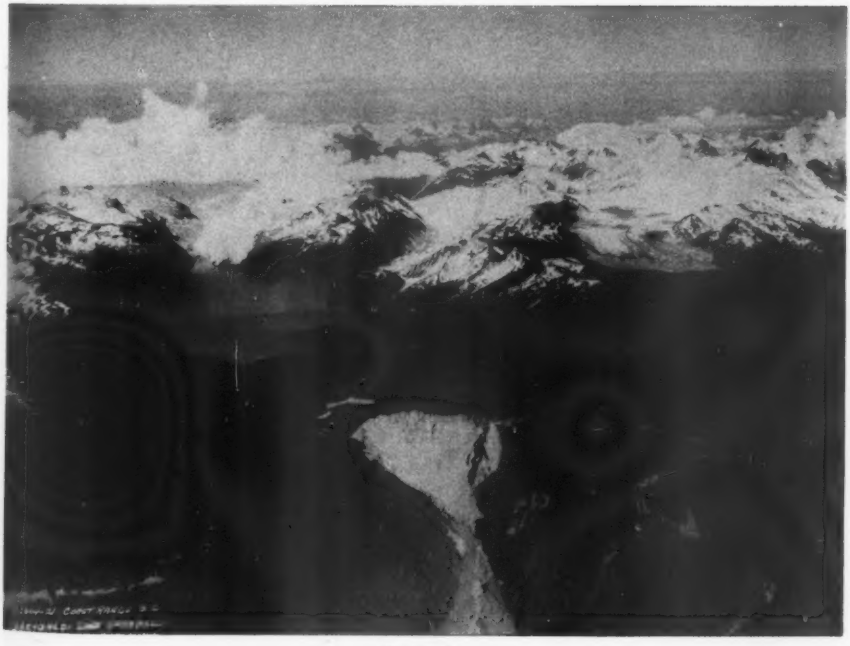


FIGURE 2

MOUNT GARIBALDI AND GARIBALDI LAKE

FIGURE 1

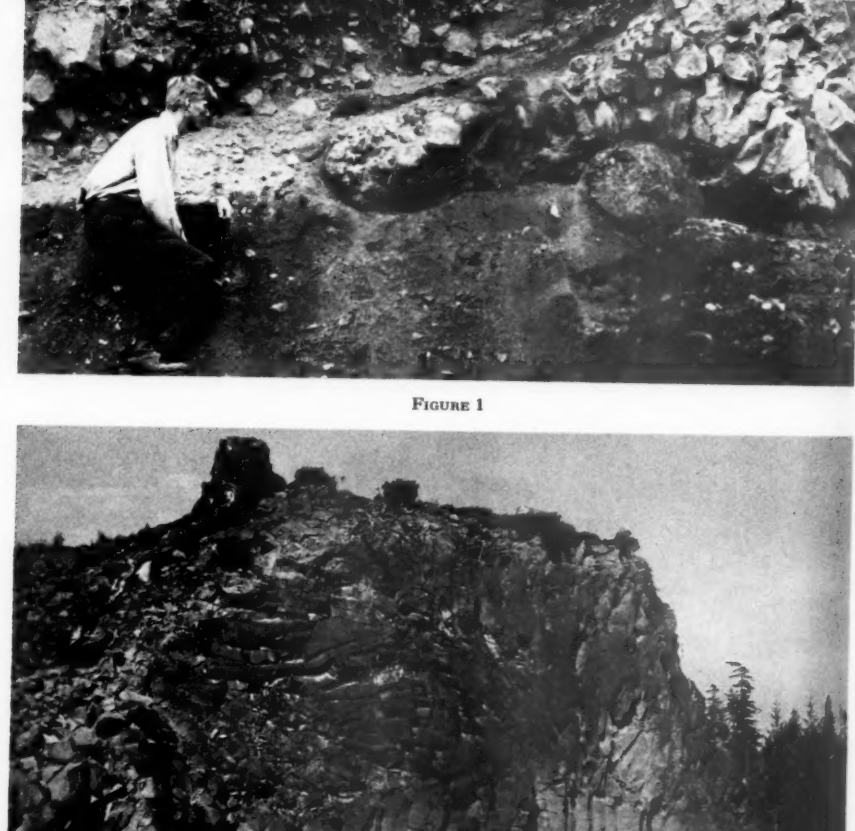


FIGURE 2

STRUCTURES IN BASALTIC FLOWS

form, to the type of bedrock, and to the time at which valley cutting was initiated. The steepest long slopes, those of Pitt Valley and the west wall of Squamish Valley, occur where most intense glacial activity is indicated by profile characteristics.

The Squamish granodiorite, the most sparsely jointed of all the rocks of the map area, underlies the steepest of the shorter slopes as well as the best examples of hanging valleys and cirques. The most constricted part of the fiord valley of Howe Sound-Squamish River is found in this granodiorite mass (Pl. 1, fig. 1). Conversely the broadest part of Cheakamus Valley is in the area of schistose quartz diorites, near Garibaldi station, where valley slopes are as low as 1200 feet per mile. Similar low valley slopes are developed on some but not all of the Upper Cretaceous sediments.

Many of the steep short valley slopes occur along streams which have been recently rejuvenated or diverted onto steeper courses by glacial or volcanic activity. Cheakamus River and Cheekye and Culliton Creeks have cut postglacial box canyons up to 500 feet deep and 1–3 miles long into declivities in their glaciated valley floors. Culliton Creek where it has been diverted by lava also passes through a spectacular canyon up to 800 feet deep and about 550 feet wide at its rim (Mathews, 1952b, p. 556). Cheekye and Mashiter Gorges also originated, following stream diversion, by volcanic activity, though before the last regional glaciation, and their greater depth and more mature valley slopes as compared with the Recent box canyons can be attributed to their greater age.

The valley erosion cycle must have occupied a large part of the Pleistocene Epoch, if not part of the Cenozoic Era as well, judging by comparison with the volumetrically insignificant results of stream cutting in about 10,000 years since the last major glaciation of this area. This cycle, on the other hand, can have occupied only a small fraction of the time required for the upland cycle.

Unmodified and Modified Constructional Surfaces

Part of the topography is a product of constructional processes, notably alluviation and vulcanism (Pl. 3). Constructional surfaces include the flat-floored flood plain of Squamish Valley, the gravel fan of Cheekye River, the century-old landslide of Rubble Creek (Mathews, 1952b, p. 559–60), the late Wisconsin terraced fanglomerates of Cheekye River (Mathews, 1952a, p. 93–95), the lava fields of Clinker Mountain (Mathews 1952b) and of Ring and Helm Creeks, and, most conspicuous of all, the volcanic pile of Mount Garibaldi (Mathews, 1952a). Only the youngest of these surfaces have escaped the general modification of erosional processes, and on the older deposits all gradations between constructional and erosional surfaces can be found. Many of the smaller Pleistocene volcanic centers are now marked only by low mounds and the oldest of the volcanic accumulations, although postdating the early gorges of the present Valley erosion cycle, have been so modified that no signs of original form can be found.

Glaciation

Although records of at least two distinct periods of glaciation have been found in the vicinity of Georgia Strait and Puget Sound, 30–200 miles south of the map-area, traces of only the last regional glaciation are recognized within the map-area.

The upper limit of glaciation is poorly marked, for in the alpine zone, where the highest traces are to be found, gently sloping rock surfaces which might show striations are rarely spared destruction by frost, and smaller erratics have become partly submerged in

PLATE 3.—MOUNT GARIBALDI AND GARIBALDI LAKE

FIGURE 1.—Mount Garibaldi from Parapet Peak showing the upper limit (*A-A'*) of the older rocks on which the volcano is built. Tantalus Range, west of Squamish River, appears in extreme right rear, Howe Sound and Georgia Strait in left rear. Blocky jointed Castle Towers trondhjemite in foreground.

FIGURE 2.—Garibaldi Lake (left center) from the northwest. The Barrier and Rubble Creek Valley in lower center. Royal Canadian Air Force photograph.

PLATE 4.—STRUCTURES IN BASALTIC FLOWS

FIGURE 1.—Terminus of basalt flow in Cheakamus Valley half a mile north of Garibaldi station.

FIGURE 2.—Cross section of a narrow basalt flow at Brandywine Falls, 3 miles north of the map-area on the Pacific Great Eastern Railway, showing the inclined columns marking the original margins of the flow. Flows in Cheakamus Valley near Garibaldi station and Daisy Lake probably possess this structure but are not as well exposed.

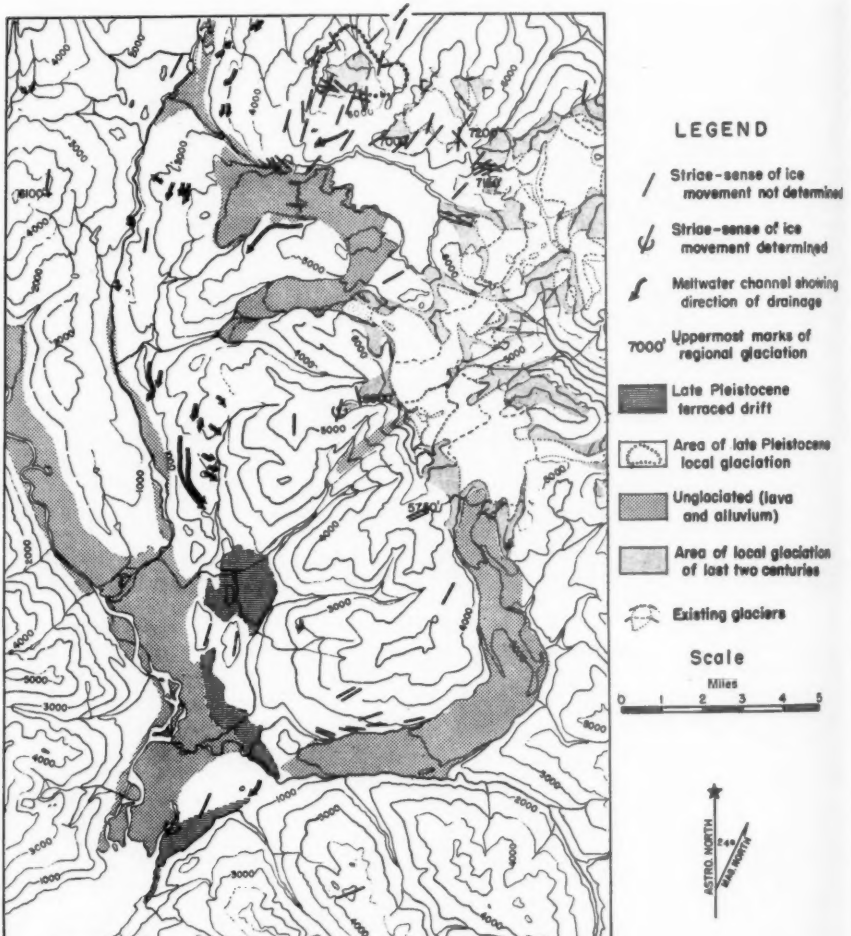

LEGEND
Strice-sense of ice movement not determined
Striae-sense of ice movement determined
Meltwater channel showing direction of drainage
7000' Uppermost marks of regional glaciation
Late Pleistocene terraced drift
Area of late Pleistocene local glaciation
Unglaciated (lava and alluvium)
Area of local glaciation of last two centuries
Existing glaciers
Scale
Miles
0 1 2 3 4 5
ASTRO. NORTH
MAG. NORTH

FIGURE 2.—GLACIAL MAP OF THE MOUNT GARIBALDI MAP-AREA

felsenmeer. Fresh erratics and striae which can be attributed to the last regional glaciation have been found in the northern part of the map area as high as 7000 feet on Panorama Ridge (Fig. 2) and the east ridge of The Black Tusk, to 7200 feet on the ridge southeast of Helm Peak. Fresh erratics and striae also occur up to 7150 feet on the west ridge of Castle Towers Mountain, but above this level are to be found extensive boulder fields developed in situ by frost wedging and weathering. These resemble the "mountain top detritus" of Dahl (1955, p. 1514–1517) who suggests that it is the product of a very long weathering cycle in unglaciated areas. At the southern edge of the map-area the ice crossed Goat Ridge at an altitude of 5500 feet but is not known to have reached higher levels (James, 1929, p. 64).

The direction of ice movement, as indicated by striae and transported boulders, averaged S. 20° W., but local deflections were common. Striae on the south slope of Round Mountain indicate westerly ice movement, though possibly these striae were cut at a late

stage of deglaciation when the ice stream was confined by the walls of Mamquam Valley. Ice moved westward and possibly northward away from the Castle Towers massif as if this area was a center of ice accumulation during the Ice Age, as it is today (Pl. 2, fig. 2). The great width and rounded form of Skookum Creek-Mamquam River Valley suggest that this trench served as an important channel for ice dispersal from the high mountains near its head. Pitt and Squamish Valleys, similarly, must have served as important channelways. Cheakamus Valley, although less obviously modified by ice, must have conducted large volumes of ice southward from the central part of the Coast Mountains. Virtually all ice from Squamish, Cheakamus, and Mamquam Valleys funneled through the fiord valley now occupied by Howe Sound.

Glaciation cannot be attributed entirely to an ice sheet which buried all but the highest peaks. Cirques, formed by local mountain glaciers, are common in the southwestern and northeastern parts of the map area. Some of these cirques would have been completely submerged during the climax of regional glaciation although others are high enough to have been occupied by alpine glaciers tributary to the ice sheet. None of the cirques in the southwestern part of the map-area, except for one on the northeast side of Sky Pilot Mountain, contains a conspicuous moraine, and these cirques must have been formed before the climax of the last glaciation. Where glaciers have been active in post-Wisconsin time, as in the northeastern part of the map-area, moraines are present, and a few are of notable size (Pl. 2, figs. 1, 2).

Meltwater channels are conspicuous in a few localities, particularly along the east wall of Cheakamus Valley. A few of these consist of narrow graded gravel plains, like the one making up the western part of the Black Tusk Meadows, half a mile north of the outlet of Garibaldi Lake. Most, however, are rock-walled trenches, a few feet to a few hundred feet deep, a few score feet to 3 miles long, and now partly filled with a recent deposit of talus fallen from the oversteepened walls and locally blocked by alluvial fans. The fresh unglaciated walls of the channels testify to their recent origin; the talus fill clearly demonstrates that the principal agent involved in their development is no longer active. Their general form and distribution, and their association with graded gravel plains points to an origin in ice-diverted meltwater streams.

Though the details of the channel floors are as a rule completely concealed by the recent fill, the direction of drainage of the channels is determinable for nearly all the larger gorges and for many of the smaller ones. This drainage is almost without exception to the south, southeast, or southwest, and reflects the southerly slope of the Wisconsin ice sheet during its declining stages.

The highest channel, marked only by a trail of stream-rounded gravel in a shallow trench, lies at an elevation of 6550 feet on the west bluff of The Black Tusk. It might be assumed that the firn line at the time of deglaciation of this bluff was still higher than the channel (Mathews, 1951b, p. 369, 371), although evidence from Sentinel Glacier, southeast of Garibaldi Lake, (Mathews, 1949, p. 284) shows that running water can be a potent agent of transportation, if not of erosion, on a glacier well above its firn line.

One channel 4.3 miles southeast of the summit of Mount Garibaldi may be of Recent origin. In 1915 it lay near the margin of one of the glaciers draining the Garibaldi névé and was then occupied by meltwater. Rampart Lake, a temporary ice-dammed body of water noted for its marked fluctuations in size, existed at this same time 1¼ miles upstream on the same ice margin, and periodic floods accompanying the draining of this lake possibly facilitated the cutting of the channel.

Quaternary Volcanic Rocks

General Statement

Garibaldi Group was the name proposed by Burwash (1914b, p. 75) for "the products of Pleistocene vulcanism...lavas and pyroclastics" in the Coast Mountains. The name was to distinguish these volcanic rocks "not only from formations of different age, but also from the Pleistocene deposits of glacial and aqueous origin, with which the volcanics are to a large extent, but not entirely, contemporaneous". Later Burwash (1918) used the term Garibaldi volcanic formation, but as his earlier name is more appropriate from the standpoints of both priority and brevity, Garibaldi Group will be used in this report. The author's work confirms Burwash's belief that the rocks in question are at least for the most part Pleistocene, but it is not yet known whether the oldest members are middle Pleistocene, early Pleistocene, or even late Cenozoic, and a possibility exists that one or two lava

flows may be of early post-Pleistocene age. One volcanic mass, The Table, which was thought by Burwash to be Miocene, is now regarded as Pleistocene (Mathews, 1951b), and a second mass, The Black Tusk, is probably also Pleistocene rather than Miocene, both are here included in the term Garibaldi Group for these Quaternary volcanic rocks.

A wide variety of rock types are present within the group: basalt, andesite, dacite, and an almost complete sequence between these rock types has been found both in the earlier and later extrusive rocks. The petrology of these rocks is already discussed in Mathews (1957).

Rocks of the Garibaldi Group are virtually confined to a belt about 10 miles wide and at least 70 miles long extending N.20°W. through and beyond the map-area (Fig. 3). The highest volcanoes, Mount Garibaldi (8787 feet above sea level), Mount Cayley (7852 feet), and Meager Mountain (8830 feet), rise 1000–2000 feet above the older rocks at their bases (Pl. 3, fig. 1) but scarcely reach the altitude of the nonvolcanic mountains 15–20 miles to the east and west (Fig. 3). The principal volcanic centers are thus confined to a broad swale in the surface linking the older rock summits of the southern Coast Mountains, although the explanation of this distribution is lacking.

The term Garibaldi Group applies to rocks having a limited spatial distribution and age range notwithstanding their wide range in composition.

Characteristics of Flows

Structures.—Flow contacts, though highly irregular in detail, commonly exhibit fair regularity on exposures more than a few tens of feet across. On precipitous cliff faces such contacts can be traced for many hundreds of feet, but on most timbered hillsides they can rarely be followed beyond the limits of a single stream canyon.

Columnar jointing is common in both acid and basic lavas. Joint columns are normal to flow contacts wherever the two features have been seen together. In two deeply dissected flows of roughly tabular form distinctly curved columns have been seen along what was once the flow margins. No deep pipe-shaped masses with curved columns such as described by Hunt (1938, p. 142–149) from the Mount Taylor lava field have been seen in the map-area. Most flows that are completely exposed from top to bottom display two layers of columns, an upper layer referred to by Tomkeieff (1940) as the entablature and a lower as the collonade. As a rule the upper layer is two or three times as thick as the lower, but locally, as in The Table (Mathews 1951c, p. 835), the upper may be as much as 20 times as thick as the lower, and in a few relatively thin sheets of lava the two tiers are of roughly equal thickness. As a rule, too, the columns of the upper tier are slender compared with those of the lower; where the contrast in thickness of upper and lower tiers (*i.e.*, in the relative lengths of columns of the two layers) is great the difference in average diameter of the columns within them is striking, and where the two tiers are of similar thickness the columns within them tend to be of the same general size. The two layers are interpreted as those parts of the flow which cooled through upper and lower surfaces respectively; greater amount of cooling through the top surface is probably responsible for the greater thickness of the upper layer, and quicker initial cooling is probably responsible for the more closely spaced joints.

Conspicuous flow bands are only locally developed. A rhyodacite intrusive mass at the eastern base of Lava Peak displays well-defined alternating layers, commonly less than a few inches thick, of glassy and of devitrified lava, paralleling the steeply dipping walls of the mass. The intrusive core of Mount Garibaldi (Mathews 1952a, p. 91) also exhibits steeply dipping flow bands, consisting of layers of slightly different color and texture, which appear to parallel its walls. Bands 1-2 feet across of alternately dull red and black dacite make up the lower part of the Clinker Mountain flow at The Barrier; the banding parallels the walls and floor of the valley into which the lava poured (Mathews 1952b, p. 559).

Orientation of microlites is very obvious in thin sections of some dacites, and hand specimens from which these sections were made exhibit silky surfaces on which this orientation can be detected. Squeezed vesicles, many of almost microscopic size, are present in some of the more acid lavas. Vesicles in the basalts are generally almost spherical and are of little or no value in determining direction of flow.

Structural types of flows.—Several types of lava flows can be distinguished on the basis of external form, internal structure, and composition.

The first type is represented in the map-

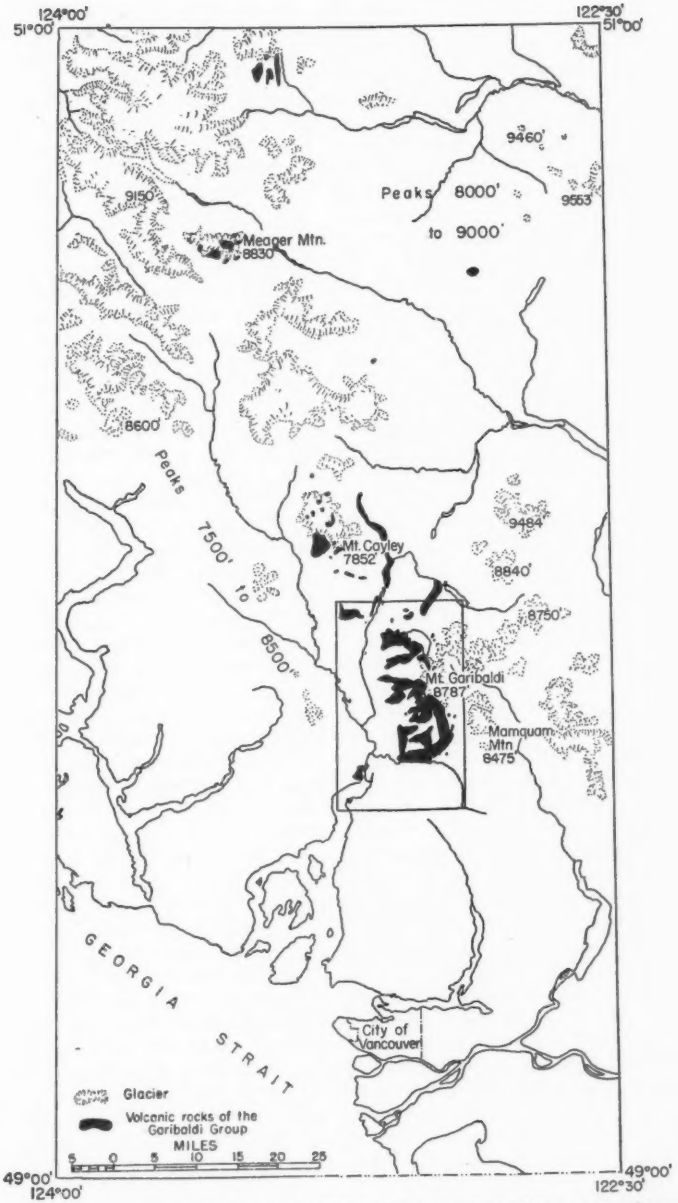


FIGURE 3.—KNOWN DISTRIBUTION OF VOLCANIC ROCKS OF THE GARIBALDI GROUP IN THE SOUTHERN COAST MOUNTAINS

area by the two major streams of lava emitted from Clinker Mountain and by the Ring Creek flow, east and south of Round Mountain. All three flows are extensive: the Ring Creek flow is 11 miles long, up to 1½ miles wide, and is estimated to be as much as 800 feet deep and to have a volume of slightly more than 1 cubic mile (Table 1); the two Clinker Moun-

tain flows are 4½ miles long, up to 1 and 1½ miles wide respectively, and combined have an estimated volume of 0.85 cubic miles. Judging from the simple levees on either side of the Ring Creek flow and from continuity of lava from base to upper surface in an exposure near its terminus, the magma comprising this voluminous flow was extruded in a single eruption. The two flows of Clinker Mountain seem also to have been the products of a single eruption (Mathews, 1952b, p. 555), but the multiple crests of the levees in the northwestern flow indicate at least three flood stages during an extended period of activity. The levees, which in the case of the Ring Creek flow rise as much as 500 feet above the axial part of the lava stream, are present in only the middle and upper parts of the flows; downstream the lava surface is flush with the walls which in most places slope outward at the angle of repose of loose debris to the older ground surface, about 200–300 feet below. The surface gradients of these flows vary from 1500 feet per mile where the lava is thin to 250 feet per mile where it is thick. Ponding of lava against the Wisconsin ice sheet has been postulated for parts of the Clinker Mountain flows which near their lower ends are deep but terminate in abnormally high and steep faces (Mathews, 1952b), but no such ponding took place at the Ring Creek flow which terminates in a sloping rubble face, about 200 feet high, similar to those along the sides of the same flow. The surface of the flows consist of assorted blocks of dacite in a layer which varies in depth from possibly a few tens of feet to as much as 200 feet. Patterns of low arcuate wrinkles on the blocky surfaces of the flows are readily apparent in air photos but not from the ground. The underlying parts of the flows, judging from the few deep exposures, consist of massive or banded lava in which is developed platy or, less commonly, widely spaced columnar jointing. The rock composing the Ring Creek and Clinker Mountain flows is dacite (Table 2, analyses 5, 6) with refractive index of fusions 1.533–1.548 (Mathews, 1951a). It is reddish near the surface but is commonly gray to black at depth. The surface lava is commonly friable because of the abundance of intercrystal cavities and minute vesicles; the lava at depth is compact.

Possible variants of this first type are relatively short, steeply dipping dacite flows occurring on the northeastern and western slopes of Mount Garibaldi and on the northeastern side of Glacier Pikes. Top and bottom surfaces of these flows dip away from their sources at angles of up to 35°. Their side walls are precipitous though it is not clear whether they were initially so or were developed later by collapse along steeply dipping joints. No great thickness of rubble covers their upper surfaces. Steeply dipping joints, dividing the rock into more or less irregular slabs, are the only internal structures recognized. Composition and petrographic characters closely resemble those of the Clinker Mountain and Ring Creek flows. Perhaps if greater volumes of lava had been extruded at any one time from the vents of Mount Garibaldi and Glacier Pikes these flows would show a much closer similarity in form and structures to the typical extrusive rocks of the first type.

The second type of flow, made up of basic andesite or basalt, is exemplified by the lavas on the floor of Cheakamus Valley north of Garibaldi station. Four occurrences of basalt south of Garibaldi station and the lava of Helm Valley, or such of it as is exposed beneath a fairly continuous cover of till, exhibit similar characteristics. The volume of basaltic lava in individual flows is small compared with the dacite extrusive rocks of Clinker Mountain and Ring Creek, although the basaltic fill in Helm Valley, made up probably of many individual flows, is estimated to have a volume of almost 0.4 cubic miles (Table 1). In the latter locality the lava extends 3½ miles from its probable source at The Cinder Cone. In Cheakamus Valley the largest flow mapped is about 2½ miles long and up to half a mile wide, although it is nowhere known to be more than 50 feet thick. The lava at all localities appears to have been rather fluid, capable of moving with low surface gradients. The Helm Creek lava has a surface slope of about 400 feet per mile, and the main flow in Cheakamus Valley a slope of about 100 feet per mile. Movement in the latter flow seems to have continued even when the lava was only a few feet thick, and the tip of the lava stream has a thickness of little more than 1 foot (Pl. 4, fig. 1). The upper surfaces of the flows are regular; although glacial erosion has in many places grooved or polished the surfaces, it does not seem to have modified their gross form. The undersides of the flows are commonly scoriaceous; a wet substratum which contributed steam as the hot lava passed over it may be responsible for the contrast between this and the upper surface. At the flow terminus mentioned above a shell of glass up to a few mm thick is present on both top and bot-

tom surfaces as well as on the walls of early fractures. Columnar jointing is almost universally present. As a rule the thickness of upper and lower tiers of joint columns are similar as are the average diameters of columns in these two layers. Vesicles tend to be almost spherical and fairly widely spaced. The rock itself is dark gray to dark brown and holocrystalline at all but the very margins of a flow. Olivine phenocrysts are generally present. In composition most of the lavas are close to the basalts and quartz basalts of the High Cascades (Williams, 1932b, p. 374 ff.) (Table 2, analyses 13, 15, 16, refractive index of fusions 1.562–1.592).

Anomalous characters in the basic flows are noted at several localities. Ellipsoidal masses, up to 20 feet across, are widespread in the southern part of Helm Creek flow and in the basal layers of the largest flow in Cheakamus Valley. A cross section of the latter flow along the railway 1.4 miles north of Garibaldi station exposes flattened ellipsoids, either pillows or ribbons of lava, a few feet thick and about 50 feet wide. Steeply sloping walls, up to 50 feet high, on either side of several flows seem anomalous in lava which elsewhere is capable of flowing in thin sheets with low surface gradients. One such steep-sided flow, west and south of Daisy Lake, though more than a mile long has an average width of only 300 feet, and its walls drop abruptly 50–100 feet to low ground on either side. A smaller flow, also long considering its width, is exposed in cross section along the railway 2½ miles north of the map-area (Pl. 4, fig. 2). Here as elsewhere, recumbent columns along the margins of the flow indicate that the steep walls are original features little or not at all modified by later erosion. The similarity of these flows to eskers in form, though of course not in internal constitution, is striking. They were developed during the waning stages of the Wisconsin ice sheet possibly by the passage of lava along tunnels or trenches thawed by heated meltwater, and if so the resemblance to eskers is more than accidental.

The third type of extrusive, steep-walled masses of dacite or andesite, are typified by the Eenostuck flow and by the west bluff of The Black Tusk; other examples are known but most have been partly destroyed by glacial and subaerial erosion. The Eenostuck flow, on Skookum Creek, is slightly less than half a mile long, has an average width at its base of about 600 feet, and has a height of fully 300 feet for the greater part of its length. Its walls are precipitous and locally overhanging, except at its lower southwestern end. At its upper end the flow is about 400 feet high and 400 feet wide at its base. Joint columns, which are well developed throughout the mass, are generally perpendicular to the present surface except near its summit where vertical columns are exposed. Evidently the principal change in this andesite mass since its origin has been the recession of its walls, most marked at its upper levels, leading to a slight decrease in their slope and to the exposure near the crest of lava which cooled through its upper rather than its lateral surfaces. The west bluff of The Black Tusk is a broad and slightly elongated mound of black dacite. Its northern, eastern, and southwestern faces are not unusually steep, but its northwestern end forms a precipitous bluff 300 feet high overlooking the long east slope of Cheakamus Valley. Here, too, columnar joints are well developed, perpendicular to the surface both of the gently dipping upper slopes and to the precipitous northwestern face. The present form of the mass evidently approximates the original shape, although an accumulation of talus at the foot of the steep face indicates some recession from its original position. One wedge-shaped gap in the northwestern face exposes the interior of the flow. Here the upper tier exhibits a beautiful fan-shaped pattern of gently curved slender joint columns sweeping outward toward the lateral margin and upward toward the top surface. This radiating aggregate rests directly on the stout vertical columns of the lower tier. Only 15 feet of this basal layer is exposed above talus, and its total thickness is unknown. One flow on the northwest side of Glacier Pikes shows this same joint pattern, though on a considerably smaller scale (Pl. 5, fig. 1). The original surface of these extrusive rocks has nowhere been found unmodified by erosion. Internal structures, other than the prismatic jointing, are not obvious; no banding has been recognized, vesicles are scarce or absent, and the change from black almost glassy lava near the outer margins of the extrusive rocks to dark gray stony or finely crystalline lava at their cores is gradual. The composition of the lava is andesite to dacite (Table 2, analysis 12, refractive index of fusions 1.542–1.562).

The form of the flows of this third type is clearly anomalous. The steep high faces assumed by the flows exhibit neither the pronounced banding nor the irregular onion-skin or fan joints so common in viscous domes or

tholoids (Williams, 1932a, p. 142–145). The andesitic composition of these lavas (Table 2, analysis 12, refractive indices of fusions 1.542–1.564) would suggest a fluidity intermediate between those of the Clinker Mountain dacite and the basalts of Cheakamus Valley. Ponding by ice of at least the steep faces of the flows (Mathews, 1951c, p. 840–841; Mathews, 1952b, p. 562–564) provides an explanation for anomalous shapes.

The fourth type of extrusive rock is represented by partly developed, or partly destroyed, coalescing dome-shaped masses of dacite, notably on Columnar Peak and on the ridge southeast of The Table. The most conspicuous single feature of these masses is the nearly perfect aggregates of radiating joint prisms, 10–50 feet long. Some of these aggregates terminate in spherical surfaces, normal to the prisms, representing either original cooling surfaces or extensive curving cross-fractures. On extrusion the dacite evidently formed a series of coalescing rounded and steep-sided domes producing, perhaps, a large-scale mammillary surface. These extrusive rocks of the Columnar Peak area consist of relatively siliceous dacite (Table 2, analysis 2, refractive indices of fusions 1.523–1.530).

A noteworthy volcanic mass which fails to conform with any of those described above is The Table, discussed in detail in a previous paper (Mathews, 1951c). This flat-topped, steep-walled, horizontally layered pile of andesitic dacite is considered to have developed by the repeated flooding of lava into a more or less cylindrical vent thawed through the Wisconsin ice sheet. Local sheets of lava coating the sides of the mass are thought to have formed where molten material locally spilled down the gap thawed between solid dacite and icy walls. No counterpart of this feature is known in the map-area save perhaps The Black Tusk (Pl. 6) which has, however, been so extensively eroded that its original form and structure are no longer determinable. The Black Tusk resembles The Table both in the composition of its lava and in the abundance of slender recumbent joint columns, as well as in the occurrence at its summit of gently dipping flows. These summit flows, however, are not flat lying and saucer-shaped as at The Table. The Black Tusk is markedly elongate in a northerly direction, parallel to the local direction of ice movement as at The Table. This elongation of The Black Tusk may be accounted for by (1) shaping of an intraglacial vent by ice movement, (2) extrusion of lava from a northward-trending fissure zone which corresponds to the direction of ice movement only coincidentally, or (3) later ice erosion.

A series of pinnacles on the west side of Squamish Valley opposite the town of Squamish also fail to conform with any of the previously described types. These consist of light gray to black dacite (Table 2, analysis 8) and rise 200 to almost 500 feet above a broader lava mass covering about 1½ square miles and mantled for the most part by dacite talus. Seen from the east under normal lighting the pinnacles are inconspicuous, but when silhouetted by clouds sweeping across the dark wooded slopes behind them they present fantastic outlines. The tallest of them, known locally as The Castle, displays a terminal spire which is particularly striking if seen from the north or south (Fig. 4). This pinnacle is, moreover, distinguished by a semicylindrical groove extending up its southern end like a flue up some huge split smokestack; the groove is bounded for its entire height by horizontally radiating dacite joint columns. The remains of a complete ring of these columns exists at the foot of the groove. Near the summit the columns now extend without interruption from the flue to the outer face of the pinnacle, but at the southwestern base the external surface is coated with a mass of agglutinated breccia. Prismatic joints extend for a few inches from the solid dacite into breccia, and indicate that the coating and the main mass cooled as a unit. More agglutinate rests on granodiorite slightly more than 100 feet east of and downslope from the base of the tower.

The origin of the pinnacles is conjectural. The dacite lava dates back only to late Wisconsin time, and erosion since that time has been wholly inadequate for the development of residual features of their height. There is no sign that they were surrounded by ice during or since development, indeed the agglutinated breccia plastered on The Castle and scattered more than 100 feet from its base suggests that this pinnacle and its immediate surroundings

PLATE 5.—STRUCTURES IN LAVA AND IN CINDER

FIGURE 1.—Margin of a flow of columnar lava, Glacier Pikes.

FIGURE 2.—Local slumping in stratified ash and lapilli, southwestern part of The Cinder Cone.

FIGURE 1

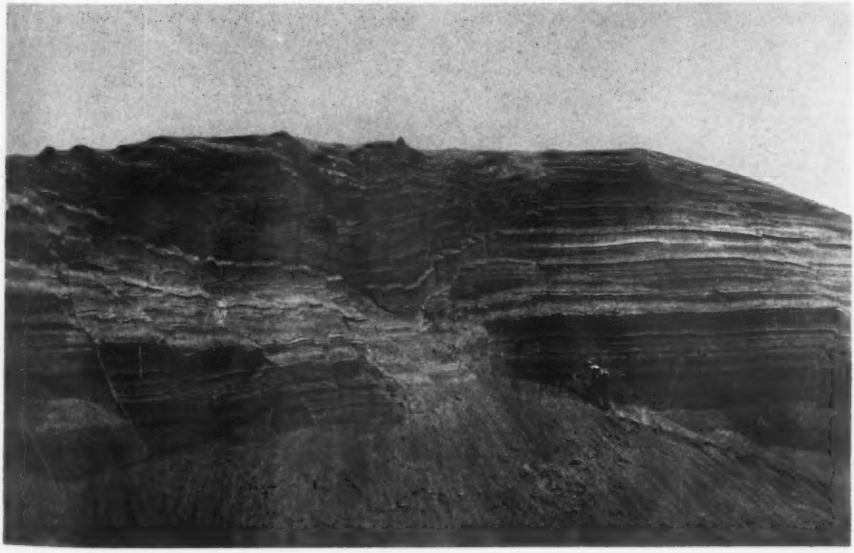


FIGURE 2

STRUCTURES IN LAVA AND IN CINDER

FIGURE 1

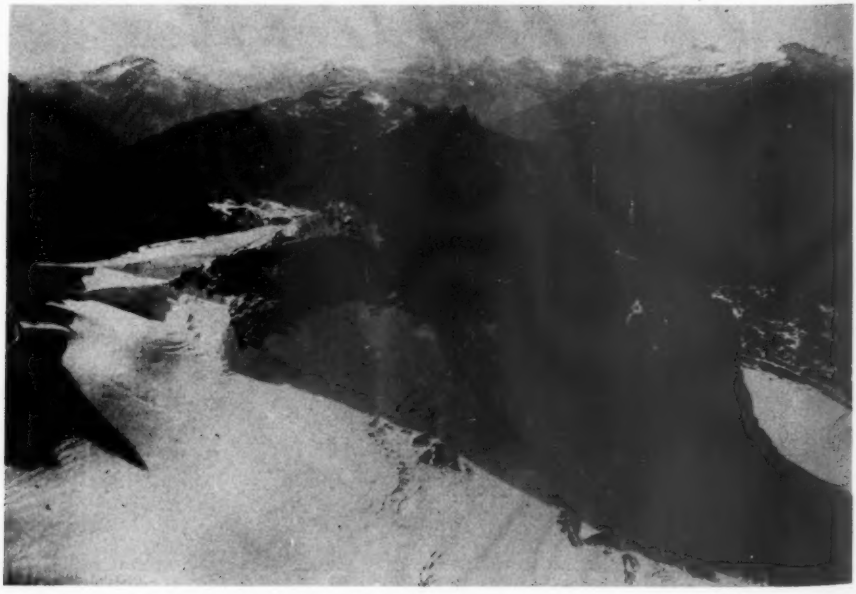


FIGURE 2

THE BLACK TUSK

were ice free while it was still hot. The Castle may have had a fluid core which was drained after partial collapse of the structure; the columnar joints may then have developed around the flue. If so it resembles in some respects Peléan spines as described by Perret (1950) who states (p. 28–29):

> "..a spine may in reality be a complete though narrow cone having a central core, perhaps still liquid or pasty, held within a cooled exterior... This was seemingly the case of the great "aiguille" of 1902.
>
> During the earlier stages of Peleean dome formation in 1930 many strange shapes were taken by the issuing material as it was forced up into view—twisted tree trunks, flat slabs with deeply grooved surfaces,.. but all were soon buried out of sight by superposition of avalanching fragments from the larger spines as the material, quasi vitreous and like unannealed glass, was shattered by the internal stresses of unequal cooling."

Many of the characteristics of Peléan spines are lacking, notably the vertical grooving and the combination of one plane face and one face smoothly curved about a horizontal axis. The striking columnar jointing of the core of The Castle is apparently not developed at Mount Pelée and, indeed, the very survival of a spine since Wisconsin time seems unusual. Thus though the pinnacles can be more closely compared to Peléan spines than to any other known volcanic form they are far from typical.

Characteristics of Pyroclastic Rocks

Pyroclastic rocks of at least two types occur in the Garibaldi Group: tuff breccias laid down by glowing avalanches during Peléan eruptions of Mount Garibaldi, and lapilli, ash, and bomb fragments ejected during Strombolian eruptions at The Cinder Cone. Some volcanic breccias elsewhere in the map area may have developed by quenching and granulation of lava injected into water, wet debris, or ice.

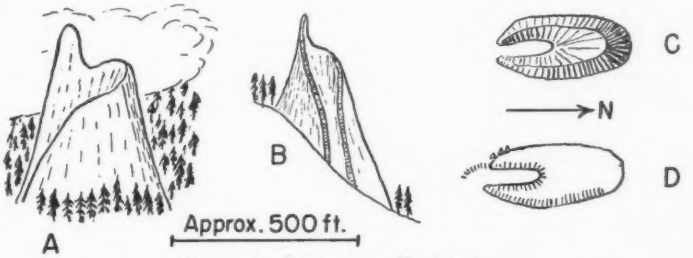


FIGURE 4.—SKETCHES OF THE CASTLE
A. The Castle from the east (from a photograph and field sketch), B. The Castle from the south (sketch based on field notes alone), C. plan, and D. horizontal section at the base of The Castle (based on field notes; in D, hatching indicates orientation of joint columns, triangles indicate breccia).

The tuff-breccias of Mount Garibaldi, described in a previous paper (Mathews, 1952a, pp. 83–85), consist of an almost completely unsorted assemblage of banded gray to reddish dacite from dust size to blocks 55 feet across. Stratification is apparent only from a distance. The deposit accumulated to a maximum thickness of about 2300 feet in the form of a broad cone, parts of which were apparently supported by tongues of the Wisconsin ice sheet. Melting of the ice permitted extensive slumping and marked modification in the form of the volcano.

The Cinder Cone, a broad mound 500 feet high, composed of ash, lapilli, and scattered ropy and breadcrusted bomb fragments, is situated at or near the source of the Helm Creek lava. It may well be related to this lava in the way that the modern volcano Paricutin in central Mexico is related to the flows which issued from its base. The Cinder Cone is distinctly asymmetrical; its northern and eastern slopes are markedly steeper and extend for lesser distances from the vent, than the southwestern slope. Moreover the bedding (Pl. 5; Fig. 2) and what are apparently relics of the original surface of the cone are distinctly irregular in the southwestern sector, and at one place they take the form of a broad saddle.

PLATE 6.—THE BLACK TUSK
FIGURE 1.—The Black Tusk from Empetrum Ridge looking southwest.
FIGURE 2.—The Black Tusk from the east. Royal Canadian Air Force photograph.

This saddle has subsequently been occupied and enlarged by the west tongue of Helm Glacier and by the stream of meltwater which issues from it.

the greater part of its site at the time of eruption.

An extensive deposit of tuff breccia fills the southernmost of a series of buried valleys

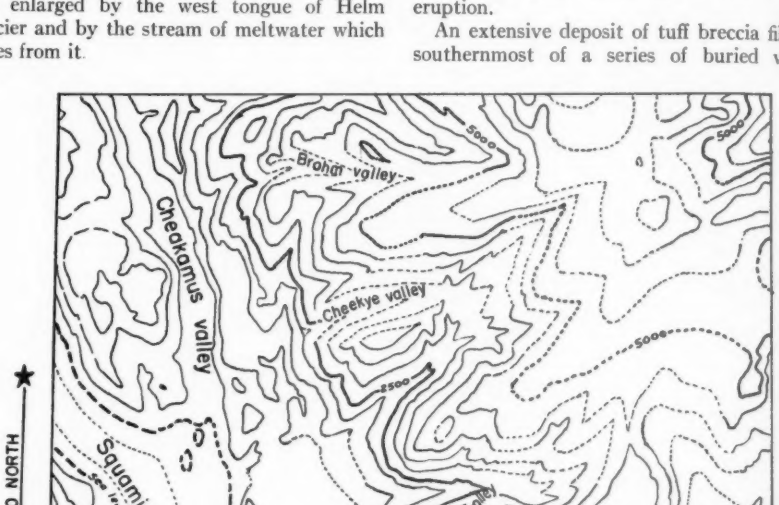


FIGURE 5.—SURFACE OF THE PRE-GARIBALDI ROCKS IN THE CENTRAL PART OF THE MOUNT GARIBALDI MAP-AREA

Contour interval 500 feet. Full lines surface now exposed and subject to erosion. Broken lines surface buried by volcanic rocks of the Garibaldi series, by alluvium, or by ice.

The asymmetry of this cone may have resulted from an inclination of the vent around which it was built. A northeasterly plunge to the vent would account for the shape and would conform with the dip of the planes of weakness provided by the stratification and schistosity in the sediments through which it passes and on which the cone was built. Possibly the southwestern part of the cone was laid down on glacial ice which, on melting, permitted local subsidence and some redistribution of the overlying ejecta. The rather perfect form of the other parts of the cone indicate, however, that ice was absent over

near Mashiter Creek (Fig. 5, Breccia Valley). The middle and upper part of the valley fill consists almost exclusively of angular fragments, generally less than 1 foot across, of buff to reddish dacite and andesite scattered through a completely unsorted matrix of comminuted plagioclase, mafic minerals, and lithic debris. Some lithic fragments show devitrification of a formerly glassy groundmass. Most of the tuff breccia is partly consolidated and some of it sufficiently so that it breaks through rather than around the more friable rock fragments. Generally, however, the matrix tends to crumble away on exposed surfaces leaving

the larger fragments standing in relief. Continuous outcrops 20 feet high show no signs of stratification, and the same type of material is found through a vertical range of more than 2000 feet. Locally fine stratified debris, generally flat lying, and coarse waterlain conglomerate are present. The base of the deposit is made up of an ill-sorted conglomerate containing subrounded pebbles and boulders of quartz diorite, greenstone, and minor dacite and andesite. Scattered boulders of the quartz diorite exceed 2 feet in diameter, and one 6 feet across was observed at the north end of the deposit. Though the material in the lower part of this valley fill may have been eroded and transported by streams or ice, the unsorted tuff breccias higher in the succession are clearly from a strictly volcanic source. Most likely they were laid down by glowing avalanches, like the tuff breccias of Mount Garibaldi, but possibly they were deposited by mudflows having their source in readily eroded pyroclastic beds.

The fill of a buried valley which drained westward from the northern side of Round Mountain (Fig. 5, Mashiter Valley) is composed in part of flat-lying masses of scoriaceous red lava 20–40 feet thick alternating with beds of red breccia generally from 10 to 30 feet thick. This breccia consists entirely of angular clinkerlike fragments of the red lava in a similar but finer-grained matrix and rare thin bands of yellowish silt. Conceivably quenching and granulation in water of some of the red lava accounts for the breccia, and the silt may represent water-deposited material laid down during times of quiescence in the volcanic activity.

Volumes of the Volcanic Rocks

A reconstruction of the ground surface buried by rocks of the Garibaldi Group (Fig. 5) can be made from information gathered in the field, and this and data on the present surface of the extrusive rocks permits reasonable estimates of the volume remaining in various volcanic masses throughout the map-area. These estimates, the refractive index, or range of indices of fusions from the rocks of which they are composed, and the relative ages as indicated by degree of dissection are given in Table 1. The distribution by volume of rocks of different composition, as determined by the refractive index of fusions and by these estimates of volumes is discussed in Mathews (1957).

Age Relationships

The members of the Garibaldi Group are clearly unconformable on all the solid rocks with which they are in contact, and they are overlain only by glacial, fluvioglacial, and alluvial debris. Most members have some topographic expression, and a few are so young that their original surfaces have only locally been scarred by stream cutting and not at all modified by glacial activity. On the other hand, a few members are sufficiently old that no trace of their original topographic expression remains, although it can be shown that they date back to earlier stages of the valley erosion cycle. No fossils have been found in any of the pyroclastic deposits, and no glacial or interglacial sediments of pre-Wisconsin age have been recognized within the map-area that might provide additional means of dating the earlier volcanic activity, but pebbles of fresh red lava, almost certainly from some member of the Garibaldi Group occur in interglacial beds at Point Grey, Vancouver. The stratigraphic relationship of one member to another with which it is in contact provides a means of establishing relative ages in a few instances, but the wide scattering of the volcanic centers renders this method of only limited application.

The oldest of the volcanic rocks of the Garibaldi Group fill a series of valleys southwest and west of Mount Garibaldi (Fig. 5). No trace of these buried valleys is preserved in the present landscape and, indeed, the volcanic rocks which filled them have themselves been so modified by subaerial and glacial erosion that their presence is not suggested by the existing topography. Nevertheless by tracing the contacts between extrusive and older rocks it is possible to locate the axes of the old valleys and to establish, within limits, their gradients. Each of these buried valleys is paralleled by a modern valley, and it appears that a former stream, displaced by a deposit of volcanic rocks, became re-established along the edge of the extrusive fill. In three of the four known examples the new valley has been cut to depths greater than that of its predecessor, but the slopes of the valley walls, the gradients of the valley bottoms, and apparently also the areas of the drainage basins are very nearly the same in the old valleys as in their younger counterparts. The similarity in valley gradients and the lack of visible folds and faults in the valley fills indicate that no major diastrophism has taken place since the valleys were buried.

The northernmost of the old buried valleys (Fig. 5, Brohm Valley) is situated in the northeastern part of Brohm Ridge. Its axis dips westward at about 660 feet per mile, and if projected at this gradient it crosses the modern course of Cheakamus River at an altitude of 1400 feet. As there is no extension of this valley, buried or otherwise, crossing the ridge west of Cheakamus River. the pre-volcanic drainage from Brohm Ridge must have joined Cheakamus River when this stream was flowing about 1200 feet above its present channel. Thus when the buried channel of Brohm Ridge became filled with lava and pyroclastic debris, Cheakamus River had cut its valley to approximately two-thirds of its present 4000-foot depth below the adjacent upland surface.

A second valley buried by volcanic rocks lies north of and parallel to Cheekye Gorge (Fig. 5, Cheekye Valley). The walls of this former valley slope inward at about 32°; its axis falls to the west about 1300 feet in 2.5 miles, and if projected onward at this same gradient it crosses the line of Cheakamus River at a point 1100 feet above sea level and 900 feet above the present stream. This channel was buried by the volcanic rocks at about the same time as the neighboring valley on Brohm Ridge.

A third buried valley, paralleling Mashiter Gorge on the south, (Fig. 5, Mashiter Valley) was cut to a depth of 1000 feet below the adjoining upland surface before being filled with basaltic lava. Since that time Mashiter Creek, in its new course, has cut a gorge as much as 2000 feet deep. This buried valley was not the first to be developed here, for it cuts discordantly across and earlier but distinctly lower

TABLE 1.—VOLUMES AND COMPOSITIONS OF MEMBERS OF THE GARIBALDI GROUP

Member	Volume (Cu. miles)	R.I. of Fused Rock*	Relative Age†
Brohm valley fill	0.10	1.528	1
Capping of Brohm Ridge	.10	1.517–1.528	1
Cheekye valley fill	.31	1.522–1.538	1
Breccia valley fill	.23	1.532–1.554	1
Mashiter valley fill	.10	1.562–1.578	1
Capping of Round Mountain	.37 (?)	1.512–1.554	1
The Table	.01	1.544–1.548	2
The Black Tusk	.01	1.538–1.543	1, 2
Lava Columnar Ridge	.21	1.523–1.530	2
Helm Creek flows	.38	1.562–1.585	2
Mount Garibaldi			
Tuff breccias	.80		
Mud slides, gravel, *etc.* in Squamish Valley	.60	1.518–1.532	2
Latest flows	.15		
Clinker Mountain			
Clinker Mountain cone	.15		
Southwest (Culliton Creek) flow	.23	1.533–1.548	2
Northwest (Rubble Creek) flow	.62		
Ring Creek lava			
The Opal Cone	.02	altered	
Ring Creek flow	1.04	1.528–1.538	2
The Castle dacites	.12	1.536–1.538	2
Cheakamus Valley basalts	.03	1.586–1.592	2
Andesites southwest of Cypress Creek	.16	1.553–1.562	?
Others	.40	1.492–1.575	
Total	6.14		

* R.I. of rhyodacite = 1.492, of dacite = 1.51–1.55, of andesite = 1.55–1.575, and of basalt = 1.575–1.592 (Mathews, 1951).

† Relative age, based on degree of dissection: 1—old, highly dissected, 2—late Pleistocene to Recent, little or not at all dissected.

channel (Fig. 5, Breccia Valley). In part because of this complication, in part because its gradient has not been precisely established, and in part because of geographic isolation from Cheakamus River and the other two buried valleys, its age relations with these two valleys is not known.

The fourth and southernmost of the buried valleys is at the western base of Round Mountain, parallel to and east of lower Mashiter Creek (Fig. 5, Breccia Valley). Unlike the other three channels it drains southward, rather than westward, and was filled with pyroclastic rocks rather than lava or a mixture of the two rock types. Its walls slope at an angle of about 25°, and its axis falls in a distance of 3 miles from an altitude of 1400 feet to less than 700 feet where it is crossed by Ring Creek. If projected but a short distance its axis meets that of Mamquam Valley only a few hundred feet above sea level and little if at all above the granitic floor of the latter valley. As there is no extension of the buried valley south of Mamquam River, it must have been tributary to Mamquam Valley. This fourth buried valley, although older than the third, was preserved by burial under a flood of volcanic rocks when Mamquam Valley had been cut to approximately its present 4000-foot depth below the upland surface.

The earliest of the volcanic activity thus took place after the last major uplift of the Coast Mountains which initiated the present (valley) erosion cycle and, indeed, after the main valleys had been cut to almost their present depths below the upland surface. Whether activity started early or late in this erosion cycle is rendered uncertain by lack of information on such factors as amount and direction of tilting associated with the uplift, and by the time and amount of glacial overdeepening.

Most of the volcanic rocks of the Garibaldi Group are sufficiently young that some trace of their presence is apparent in the existing topography. In places only a low eminence remains at the site of an accumulation of volcanic rocks; in others much of the original form remains; and in still others parts of the original upper surfaces persist.

An approximate dating of the late Wisconsin volcanic activity is possible if there can be found some indication of the thickness of the waning ice sheet at the time of eruption. Major activity at Mount Garibaldi, for example, (Mathews, 1952a) is believed to have occurred when the ice surface lay at about the present 4400-foot level, about 2000 feet below the maximum height reached by the ice at this vicinity. The Clinker Mountain flows were evidently extruded (Mathews, 1952b, p. 561–562) when the land had become ice free above the 4000-foot level, almost 3000 feet below the maximum height reached by the ice at this more northern latitude. The dacite mass west of Squamish is unglaciated down to an altitude of a few hundred feet above sea level, but its lower slopes do bear ice-deposited erratics. This extrusive rock, therefore, dates back to a very late stage in the shrinking of the Wisconsin ice sheet, to the time when the ice surface had fallen roughly 5000 feet from its flood stage. Doubtless the southward slope of the ice surface, roughly 75 feet per mile at the time of the Wisconsin climax, increased as recession continued because of more rapid melting at lower levels. A 5000-foot drop of the ice surface in the southern part of the map-area might involve no more time than a fall of 3000–4000 feet in the northern part. Thus in the absence of data on the slope of the ice surface at various stages of retreat neither the elevation of the ice relative to present sea level nor the elevation relative to flood levels constitute precise indexes of time.

Only two flows are known in the map-area which show no signs of contact with glacial ice. One of these poured down the western slope of Mount Garibaldi after withdrawal of ice from the base of the mountain permitted extensive subsidence of the tuff breccia cone (Mathews 1952a, p. 102). Partial destruction of this flow by additional sapping of the tuff breccia on which it lies suggests it was extruded not long after withdrawal of the ice. Though the Ring Creek flow also shows no sign of either glacial erosion or glacial ponding it probably came to rest soon after the ice sheet had disappeared from the vicinity of its terminus. The tip of the flow entered a gap between two low hills at the mouth of Mamquam Valley, a gap apparently previously occupied by Mamquam River, and very nearly reached a fluvioglacial terrace on the eastern side of Squamish Valley (Mathews 1952a, p. 95). Mamquam River was diverted by the lava to its present site before wasting of ice in Squamish Valley permitted it to flow across and dissect this terrace. This flow is probably of late Wisconsin age, little if at all younger than the partly glaciated flows farther north.

The distribution by age of the later members of the Garibaldi group raises an interesting point. During the waning stages of the Wisconsin ice sheet a relatively large volume of magma,

TABLE 2.—CHEMICAL ANALYSES OF ROCKS OF
Analyses by W. H. Herds man, Gla

	1	2	3	4	5	6	7	8
SiO_2	74.84	65.24	64.40	63.94	63.08	61.07	60.92	60.53
Al_2O_3	14.66	17.92	17.72	16.48	18.06	18.46	19.74	17.42
Fe_2O_3	.37	1.28	1.61	3.38	1.62	1.90	2.32	2.82
FeO	1.21	2.57	2.99	1.29	3.06	3.63	2.98	2.73
TiO_2	.12	.40	.60	.46	.58	.56	.66	1.04
MnO	.07	Trace	Trace	.23	Trace	.14	.18	.12
CaO	1.36	4.92	5.04	5.38	5.63	5.78	5.14	5.60
MgO	.34	1.94	2.15	2.53	2.59	3.22	2.13	2.74
K_2O	2.38	1.65	1.48	1.42	1.22	.99	1.20	1.13
Na_2O	4.49	3.46	3.26	3.64	3.72	4.23	3.63	3.98
H_2O, −105°C	Nil	.14	.12	.16	Nil	Nil	.24	.48
H_2O, +105°C	.25	.63	.24	.98	.08	.07	.83	.92
CO_2	Nil	Nil	Nil	Nil	Nil	Nil	Nil	Nil
P_2O_5	Trace	.11	.16	.09	.07	.16	.22	.21
	100.09	100.26	99.77	99.98	99.71	100.21	100.19	99.72
								NORMATIVE
Quartz	35.38	24.33	24.80	22.67	19.55	13.95	20.39	17.06
Orthoclase	14.06	9.75	8.74	8.39	7.21	5.85	7.09	6.68
Albite	37.97	29.26	27.57	30.78	31.46	35.77	30.70	33.65
Anorthite	6.75	23.69	23.96	24.44	27.48	27.62	24.06	26.33
Corundum	2.22	1.76	1.98	.....	.55	.31	3.66	.....
Diopside	.....	.....	.....	1.29	.....	.....	.....	.06
Hypersthene	2.70	7.83	8.52	5.70	9.77	12.45	8.10	7.99
Olivine	.....	.....	.....	.....	.....	.....	.....	.....
Magnetite	.54	1.86	2.33	3.58	2.35	2.76	3.36	4.09
Hematite	.....	.....	.....	.91	.....	.....	.....	.....
Ilmenite	.23	.76	1.14	.87	1.10	1.06	1.25	1.97
Apatite	Trace	.25	.37	.21	.16	.37	.51	.49
H_2O+	.25	.63	.24	.98	.08	.07	.83	.92
H_2O-	.....	.14	.12	.16	.....	.....	.24	.48
Total	100.09	100.26	99.77	99.98	99.71	100.21	100.19	99.72
% An in plag.	14.3	43.4	45.0	42.8	45.0	42.2	42.5	42.5
% MgO in mafics	37.5	67.9	67.8	100.0	71.8	100.	71.3	88.3
R.I. of fusion	1.4925	1.5295	1.5275	1.5315	1.5345	1.5375	1.5340	1.5365

1. Devitrified rhyodacite, eastern base of Lava Peak.
2. Dacite breccia, south face of Columnar Peak.
3. Dacite flow, southwest end of Brohm Ridge.
4. Dacite debris from glowing avalanche, Diamond Head.
5. Dacite flow, Ring Creek lava, from Mamquam River near the terminus of the flow.
6. Dacite flow from Clinker Mountain, at The Barrier.
7. Dacite dome (?), 0.8 miles northwest of Garibaldi station.
8. Dacite spine (?), The Castle, west of Squamish.

roughly 4 cubic miles, was extruded; since the ice sheet disappeared little or no magma has been poured out. During the waning stages of the ice the coastal region has risen isostatically perhaps as much as 1000 feet; since that time it has remained stable. Could the volcanic activity be genetically related to this isostatic uplift, and now that isostatic equilibrium has apparently been restored is magmatic activity, if any persists, being confined to the depths?

ROCKS OF THE GARIBALDI GROUP
man, Glasgow, Scotland

	9	10	11	12	13	14	15	16
SiO_2	59.65	57.94	57.48	55.97	54.62	52.16	49.68	48.92
Al_2O_3	19.14	19.07	18.97	16.52	19.27	16.92	16.39	18.33
Fe_2O_3	1.51	2.58	2.31	1.60	1.13	7.07	1.83	2.10
FeO	3.97	3.64	3.88	5.18	7.16	1.12	9.77	9.09
TiO_2	.76	.82	.90	1.02	1.12	.88	1.34	1.88
MnO	.14	.06	.16	.14	.09	.24	.26	.25
CaO	6.12	6.83	6.62	8.06	6.96	9.78	8.84	8.92
MgO	3.05	3.66	3.18	5.82	4.93	7.24	7.62	5.87
K_2O	1.24	.78	1.04	1.24	.64	.73	.54	.64
Na_2O	3.77	3.55	4.12	3.25	3.88	3.36	2.72	3.42
H_2O, −105°C	.03	.14	.12	.14	.18	.12	.38	.13
H_2O, +105°C	.56	.66	.76	.82	.10	.18	.52	.33
CO_2	Nil	Nil	Nil	Nil	Nil	Nil	Nil	Nil
P_2O_5	.09	.22	.28	.16	.14	.28	.23	.18
	100.03	99.95	99.82	99.92	100.22	100.08	100.12	100.06
COMPOSITION								
Quartz	13.30	13.35	9.76	6.39	3.04	2.71	.....	.....
Orthoclase	7.32	4.61	6.14	7.32	3.78	4.31	3.19	3.78
Albite	31.88	30.02	34.84	27.48	32.81	28.41	23.00	28.92
Anorthite	29.78	32.44	30.21	26.84	33.28	28.94	30.92	32.78
Corundum	.68	.50	.....	.....	.....	.....	.....	.....
Diopside	.....	.....	.65	9.77	.27	13.82	9.38	8.46
Hypersthene	12.64	12.42	11.63	16.53	22.66	11.62	21.38	7.24
Olivine	.....	.....	.....	.....	.....	.....	5.63	11.39
Magnetite	2.19	3.74	3.35	2.32	1.65	1.85	2.65	3.04
Hematite	.....	.....	.....	.....	.....	5.80	.....	.....
Ilmenite	1.44	1.56	1.71	1.94	2.13	1.67	2.54	3.57
Apatite	.21	.51	.65	.37	.32	.65	.38	.42
H_2O+	.56	.66	.76	.82	.10	.18	.52	.33
H_2O-	.03	.14	.12	.14	.18	.12	.38	.13
Total	100.03	99.95	99.82	99.92	100.22	100.08	100.12	100.06
% An in plag.	46.7	50.5	45.0	45.5	48.8	49.0	55.9	51.5
% MgO in mafics	66.3	78.3	72.1	73.8	60.5	100.0	62.9	61.0
R.I. of fusion	1.5435	1.5520	1.5465	1.5640	1.5625	1.5775	1.5835	1.5845

9. Dacite flow, near summit of The Black Tusk.
10. Dacite dike in breccia, altitude 2,800 feet, southwest slope of Round Mountain.
11. Andesitic dacite, The Table.
12. Andesite, Eenostuck flow, Skookum Creek.
13. Andesite bomb, The Cinder Cone.
14. Oxidized 'basalt', altitude 2,600 feet, west slope of Round Mountain (buried valley of Mashiter Creek).
15. Basalt flow, west of Daisy Lake.
16. Basalt flow, ½ mile northwest of The Cinder Cone.

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PETROLOGY AND ORIGIN OF THE POWAY CONGLOMERATE, SAN DIEGO COUNTY, CALIFORNIA

BY GEORGE J. BELLEMIN AND RICHARD MERRIAM

ABSTRACT

The Poway Conglomerate is an Eocene formation, covers approximately 125 square miles in western San Diego County, California, ranges in thickness up to about 1000 feet, is nearly horizontal, and in most places lies unconformably on the Cretaceous Peninsular Range batholith and accompanying metamorphic rocks.

Most of the formation consists of pebble and cobble conglomerate, although boulder conglomerate and sand lenses occur. More than 70 per cent of the clasts are soda rhyolite porphyry and tuff. Aplites, granophyres, and metamorphic rocks are other important constitutents.

The volcanic rocks and granophyres have not been found in place, and the writers suggest that they were shallow intrusives into and effusives onto the roof of the batholith and have been removed by erosion. The resistant rhyolitic rocks withstood weathering and abrasive action and were deposited during the Eocene as the Poway Conglomerate. A textural, mineralogical, and chemical gradation from batholithic rocks to rhyolitic rocks of the clasts is taken as evidence supporting this theory. Deposition was largely continental but also marine in part.

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Introduction and Acknowledgments

The primary purpose of this study is to map and study the outcrops of the Poway Conglomerate, to make a petrographic study of the pebbles, and to determine the relative abundance of each rock type. The secondary objective is to offer a hypothesis for the origin of the conglomerate.

The area of exposed Poway Conglomerate studied is in San Diego County, in southern California on the west side of the Peninsular Ranges (Fig. 1). The main body is a dissected oval area of more than 125 square miles immediately north of the San Diego River. Other exposures are found to the southeast of the San Diego River around La Mesa and El Cajon. North of Poway Valley and the town of Poway there is a small area of conglomerate about 2 miles long, and to the east ridges of Poway Conglomerate cap several hills south and east of Ramona. The conglomerate undoubtedly also extends south and southeast under a cover of later sediments. The formation is composed chiefly of well-rounded cobbles with sparsely distributed sand lenses and a few bouldery phases. The latter contain boulders ranging up to 10–15 feet and locally appear to be younger channel fillings. The entire section of sediments is exceptionally massive, and structures such as stratification and cross-bedding are extremely vague or lacking. The dip appears to be essentially horizontal, although a low southeasterly dip may prevail in the western portion.

The work on this report was started in the summer of 1951 by Bellemin. His purpose was to make a thorough study of the character of the Poway Conglomerate pebbles. After this preliminary work Merriam joined in the study, which was expanded to include the petrographic relation of the Poway pebbles to the known basement rocks of the area with the intention of determining the origin of the formation. In 1954 a research grant from the Penrose Bequest of The Geological Society of America made it possible to obtain chemical analyses of the Poway pebbles.

Grateful appreciation is expressed to The Geological Society of America for the grant, without which the study could not have been completed, to Dr. A. O. Woodford, whose criticism has been invaluable in this study, and to R. D. Terry, who drafted the chart and geological map.

History of Poway Conglomerate Studies

The first geological mention of the Poway Conglomerate appears in the reports of Blake (1856) on the exploration and survey of a railroad route from the Mississippi River to the Pacific Ocean. During the ensuing period of more than 50 years the rocks were referred to only twice. Both references were made in reports issued by the California Mining Bureau (Hanks, 1886; Fairbanks, 1893). Neither report contributes much information on the nature of these conglomerates.

The first real geologic work done in the San Diego area was a geology and ground-water study by Ellis and Lee (1919). The name Poway was applied to the thick conglomerate beds of the area which were described in general terms but not dated. Two papers by Hanna (1926a; 1926b) on the geology of the La Jolla quadrangle give the stratigraphy of the area and describe the Poway Conglomerate, correlating it tentatively with the Tejon Formation (upper Eocene). Further work by Dusenbury (1932) and Cushman and Dusenbury (1934) on the Foraminifera confirmed this date. Later, a vertebrate fauna was found in the Poway by Stock (1937; 1938); this indicated an Eocene age which agreed in general with that suggested by the Foraminifera. More recently Hertlein and Grant (1939; 1944; 1954) have described the Poway Conglomerate in connection with studies of the San Diego region.

The observations and conclusions of all these authors may be summarized as follows:

(1) Most of the Poway outcrops are north of the San Diego River, between the San Diego River Canyon and Poway Valley; small patches appear in the surrounding areas, particularly to the northeast.

(2) The formation is not of a uniform thickness. The greatest thickness is probably not more than 1000 feet.

(3) The pebbles are large and consist mainly of red porphyry, probably andesite, of unknown source.

(4) The formation contains both marine and continental facies with a probable interfingering of beds very much like parts of the Sespe Formation in Santa Barbara County, California, 200 miles northwest of the Poway area.

(5) It is of Upper Eocene age and equivalent to the Tejon in California, Upper Claiborne of the Gulf Coast, and approximately equivalent to the Bridger Upper Middle Eocene.

Regional Geology and Stratigraphy of the Area

The Poway Conglomerate lies on the west slope of the Peninsular Range, the chain of

mountains extending north and south as the backbone of southernmost California and northern Baja California. The range is a geological unit consisting of the composite Peninsular Range batholith and associated metamorphic rocks. The relative proportions of these two groups differ from place to place, but in general the batholithic rocks greatly predominate.

The metamorphic rocks are complex and of variable composition. A group of metasediments represented by slates, sericite schist, quartz-mica schist, quartzite, and other types is widespread. Metavolcanic rocks termed the Black Mountain Volcanics by Hanna (1926a) and Santiago Peak Volcanics by Larsen (1948) compose another group. They are chiefly mildly metamorphosed flows and pyroclastics of andesite and quartz latite with some rhyolite and basalt. Most of these are dense, fine-grained, gray to black rocks with aphanitic groundmasses. The rhyolitic members are rare and have been described as flows or sills. They are white to greenish white and have abundant phenocrysts of quartz and plagioclase (andesine to oligoclase). Small intrusive bodies of related rocks occur with the volcanic rocks. Most are fine-grained porphyritic granodiorite.

In the Corona and Elsinore quadrangles, about 50 miles north of the Poway outcrops, an area of 20 square miles is underlain by the Temescal Wash quartz latite porphyry. This is a dark-gray to black rock which has a dense aphanitic groundmass and prominent phenocrysts of quartz and andesine. It is considered to be a pre-batholithic rock.

Except for the cover of alluvial patches, small areas of late Tertiary volcanic rocks, and the Poway and related sediments in the west, the batholith is well exposed. Individual intrusions within the batholith range from gabbro to granite. These intrusions form a group showing systematic variations in mineral composition, chemical composition, and sequence of emplacement and are therefore believed to have formed from a primary gabbroic magma by crystal differentiation possibly modified by limited assimilation. Of the known batholith more than half is tonalite, one-third granodiorite, and 2 per cent granite. Extensive unmapped eastern parts are high-silica, low-potassium tonalites and granodiorites.

Texturally all members of the batholith are generally medium-grained granitic, and only a few rocks, principally granodiorites, are fine-grained granitic and nearly aplitic. Granophyric texture occurs in only one small body near Corona (50 miles north of Poway Valley), and aplitic rocks, although widespread, are relatively very minor in volume.

Several parts of the batholith have been repeatedly dated as about 100 million years old, by several element-transformation methods. The geologic age is, therefore, probably Cretaceous (Larsen *et al.*, 1954; Jaffe *et al.*, 1955).

All rocks of the batholith are notable for the unmetamorphosed condition of minerals and textures. Except for minor deuteric changes affecting some of the earliest ferromagnesian minerals and incipient alteration of feldspars in the granodiorites and granites, the rocks are unaffected.

The eastern, topographically higher portions of the Poway Conglomerate lie directly on an irregular bedrock surface, but to the west the conglomerate lies on older marine sediments. The stratigraphy of these sediments has been summarized by Hertlein and Grant (1954). The oldest exposed sediments are Upper Cretaceous, in part at least Campanian gray to brown shales and well-cemented sandstones, which have an approximate thickness of 500 feet. However, deep wells penetrate a conglomerate which may be older, possibly Trabuco (Upper Cretaceous) in age.

Unconformably overlying the Cretaceous is the Eocene La Jolla Formation, composed of three conformable members as follows: lower, Delmar Sand 100 feet; middle, Torrey Sand, 20 to 200 feet; upper, Rose Canyon Shale, 300 feet. Where the Poway rests on the La Jolla Formation the contact is in most places conformable but locally unconformable.

The Poway is locally covered unconformably by more than 1000 feet of soft-gray to yellow marine sands of Pliocene age termed the San Diego Formation. This, in turn, is overlain by the Upper Pliocene to Pleistocene Sweitzer Formation composed of reddish-brown conglomerate with sandy streaks. The Sweitzer is less than 30 feet thick, caps marine terraces in the vicinity of San Diego, and consists largely of reworked Poway pebbles and cobbles.

Pleistocene marine-terrace deposits 20–30 feet thick are common near the coast.

Stratigraphic Column of the San Diego Area

Sweitzer Formation 20 feet	Interbedded reddish-brown gravels and beach sands	Pliocene and Pleistocene
	Unconformity	
San Diego Formation 1000 feet	Coarse and fine-grained yellow to brown sandstones with stringers of gravel	Pliocene

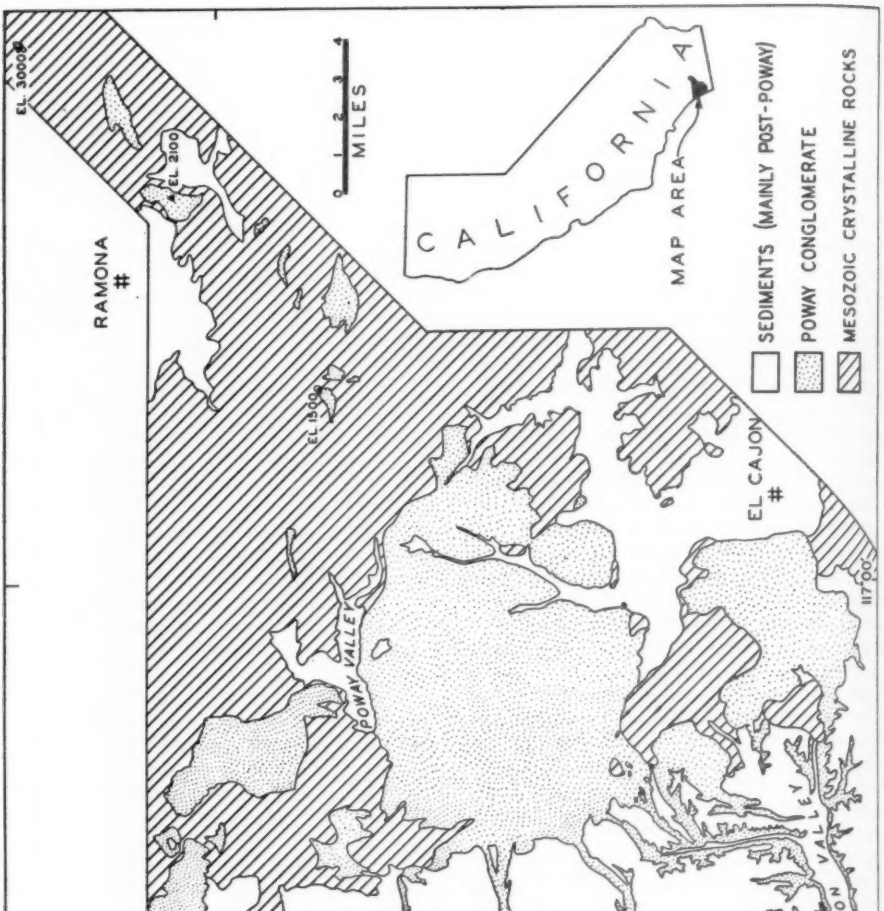

FIGURE 1.—Geological Map Showing Distribution of the Poway Conglomerate

	Unconformity	
Poway Conglomerate 1000 feet	Massive conglomerate of well-rounded pebbles and some boulders, consisting largely of rhyolitic tuff and porphyry; interbedded and cross-bedded with sand lenses of coarse sand and fine silty clay; upper zones contain caliche in many places.	
	Unconformity (?)	
Rose Canyon Shale 300 feet	Massive and laminated brownish siltstone with some members of yellow sand and stringers of gravel	Eocene
	Unconformity (?)	
Torrey Sand 20-200 feet	Massive, coarse, arkosic, white sand, cross-bedded in many places	
Del Mar Sand 100 feet	Greenish-gray, pink, purplish-red sand and shale	
	Unconformity	
Chico (?) Formation 1000-2000 feet	Brown sandstones, shales with some bluish-gray mudstones and gray or black fossiliferous shale	Cretaceous
	Unconformity	
Trabuco (?) conglomerates	Reddish-brown sandstones and conglomerates	
	Unconformity	
Batholith of Southern California	A composite, quartz-plutonite batholith consisting of about 20 distinct rock types	
	Intrusive contact	
Black Mountain Volcanics (Santiago Peak Series?) 2000 feet	Extrusive rocks, flows that are primarily andesitic and pyroclastic rocks, metaconglomerates, sandstones, and shales	Jurassic (?)

FIGURE 1.—Geological Map Showing Distribution of the Poway Conglomerate

Poway Conglomerate Collections

Method of Study

The samples for pebble counts were obtained in about the same way at each locality. The usual sample consisted of a complete sample of the conglomerate, matrix and pebbles, with not less than 100 pebbles. Fifteen such samples of the Poway Conglomerate were collected; of these only 3 had less than 100 pebbles, and 4 had more than 200 pebbles.

The sample localities (Fig. 2) were selected as typical of the local section. A small hole (10 by 10 by 8 inches) was then dug, and all the material from the hole was put into a sack or box. Where the conglomerate was very well cemented the collection was made from a larger surface area by picking at the surface and collecting all the pebbles knocked loose. Before leaving a locality a check was made in the general area to see if any significant features or clasts of unusual composition were apparent in the outcrop. At no time were random samples collected for a pebble count.

The sample was soaked in water to loosen the pebbles from the matrix; then it was passed through a screen with a quarter-inch mesh. The material that passed through the screen was considered as the matrix. The longest diameters of all clasts on the screen were then measured with a steel ruler or 6-inch calipers. The small pebbles, with lengths between 1 inch and a quarter of an inch, were checked for significant features but otherwise not included in any further study.

All the pebbles more than 1 inch long were broken and studied with a hand lens. The hand-lens determinations were adequate to distinguish most of the rock species. However, in the rhyolitic rocks albite was rarely distinguishable, primarily because the very fine discontinuous twinning was obscured in most rocks by alteration of the feldspar. From this study the rock species were determined, and typical specimens for thin sectioning were selected.

The Poway Conglomerate collections described are numbered by collection localities in the sequence in which they were taken and recorded. They are numbered PC (Poway Conglomerate) 1 to 19. At locality PC-2 no collection was made, and at PC-10, 11, and 12 there was some question of whether the material was Quaternary alluvium or Poway, and therefore these four localities were not included in the study. The collecting localities are shown on the accompanying map. Table 1 summarizes the pebble count for each locality.

Description of Poway Conglomerate Collections

POWAY CONGLOMERATE SAMPLE PC-1: The first sample was taken on the south side of the San Diego River in an area mapped as Poway Conglomerate by Hanna (1926a). In this area Poway Conglomerate is present on both sides of the San Diego River. The sample was collected in a gully 20 feet off the west side of Fairmount Road 1 mile north of El Cajon Boulevard. One long stringer of sand shows that the beds are practically horizontal.

The size distribution of the clasts indicates a maximum at 1 or 2 inches in diameter. The largest boulder is about a foot long. Nearly all pebbles and larger clasts are well rounded and ellipsoidal in shape.

The matrix consists of angular grains, ranging in size from grit to fine sand. Quartz and feldspar are the most abundant minerals, about

equally divided, and amounting to 95 per cent of the matrix. Most of the remaining 5 per cent is weathered biotite. Most of the quartz grains are clear, and many have a roughly hexagonal shape, which suggests that their probable source was a phenocryst weathered free from a porphyry.

1½ inches in diameter and a larger piece of a tooth resembling a molar imbedded in bone. Both of these were in a sandy layer and were weathered to the point where they crumbled to the touch.

POWAY CONGLOMERATE COLLECTION PC-3: Collection PC-3 was obtained beneath a Pleisto-

TABLE 1.—SUMMARY OF POWAY CONGLOMERATE PEBBLE COUNT

	PC-1	PC-3	PC-4	PC-5 & 9	PC-6	PC-7	PC-8	PC-13	PC-14	PC-15	PC-16	PC-17	PC-18	PC-19
Rhyolitic porphyry or tuff	156	101	103	168	92	102	193	178	43	56	88	86	130	170
Felsitic rocks	11	9	77	0	4	8	8	4	15	8	5	6	10	15
Aplitic rocks	7	7	1	1	2	3	3	12	0	1	4	1	1	9
Granitic rocks	30	0	0	16	6	0	3	8	1	1	6	0	3	7
Quartzite	12	24	8	2	2	44	15	3	1	0	4	7	2	2
Andesite porphyry	10	0	4	8	0	0	10	2	0	0	3	0	0	0
Miscellaneous and unknown	10	12	20	37	14	24	39	24	3	6	1	9	7	29
Total	236	153	153	232	120	181	271	231	63	72	111	109	153	232

The pebbles are 66 per cent rhyolitic rock; most other samples are even higher in this constituent. The most interesting specimens of the granitic group were six pebbles that had a roughly concentric structure and uneven texture. These may be parts of orbules or inclusions. Unfortunately all the orbicular specimens were weathered and crumbled away. Most of quartzite is light brown. The felsitic material is light and dark, and the darker material resembles basalt. The aplitic pebbles were classed as such on the basis of the sugary texture. Plagioclase is abundant enough in some to permit classifying them as granodiorite aplites.

Table 2 gives a pebble count on small material from the same locality. Even though the texture could not be determined in each specimen, the porphyry (or welded tuff) is dominant, but it has diminished with respect to the other rock types. The greater abundance of siliceous "porphyries" among the larger clasts is perhaps due to the great resistance of the rhyolite group compared to a more easily fractured rock which might shatter and produce many small fragments, while the rhyolite remains whole. This also shows that the larger clasts, more than 1 inch, are reliable indicators of the character of of the whole conglomerate.

About 75 feet up canyon on the same side of the gully some vertebrate fossil material was found. It consisted of a tusklike tooth about

cene terrace on the north side of Carroll Canyon north of Miramar, on the former course of U. S. Highway 395. The vertical exposure of conglomerate is well cemented; blocks of conglomerate that break away, because of undercutting, hold together in large blocks. The thin red adobe Pleistocene soil penetrates into the conglomerate and colors the top 2 or 3 feet. The beds are apparently roughly horizontal.

The size distribution of the pebbles is approximately the same as at PC-1, but the matrix is different. There is an abrupt break in the sorting of grains less than a quarter of an inch in size with a silty-sandy clay as the matrix.

The rhyolite porphyry and tuff described in PC-1 is again found here in almost the same percentage of the total. The felsitic rocks are about as abundant as in PC-1. One specimen shows flow structure. The aplites are about the same as in the previous collection. A specimen that resembles a lamprophyre (dark gray) was included in this group.

POWAY CONGLOMERATE COLLECTION PC-4: Sample PC-4 was collected from the hill at the top of the Poway grade on the former course of U. S. Highway 395. The hilltop (elevation 1022 feet) on the east side of the road is part of the ridge that is cut by the road. The conglomerate is well cemented in a sandy clay matrix. The bedding is roughly horizontal, based on some sand lenses visible in the road cut.

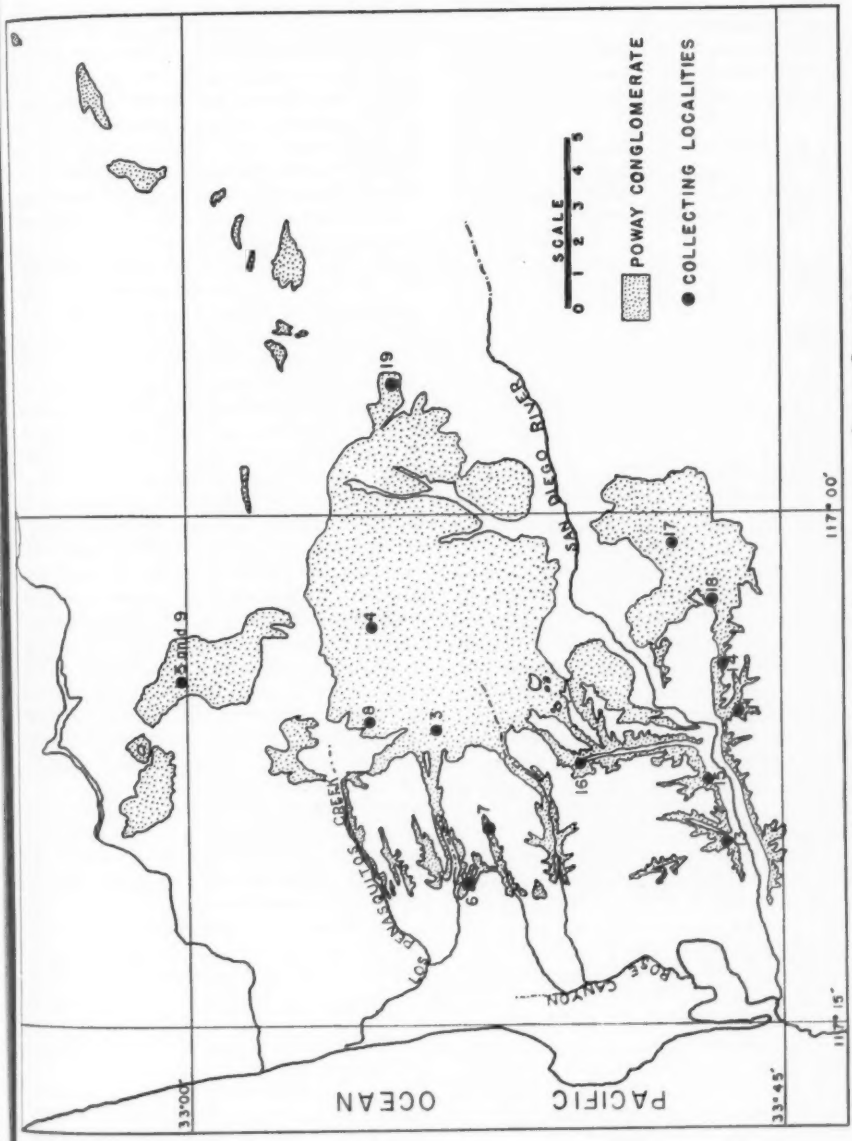

FIGURE 2.—MAP OF COLLECTING LOCALITIES OF THE POWAY CONGLOMERATE

Introduction and Acknowledgments

The primary purpose of this study is to map and study the outcrops of the Poway Conglomerate, to make a petrographic study of the pebbles, and to determine the relative abundance of each rock type. The secondary objective is to offer a hypothesis for the origin of the conglomerate.

The area of exposed Poway Conglomerate studied is in San Diego County, in southern California on the west side of the Peninsular Ranges (Fig. 1). The main body is a dissected oval area of more than 125 square miles immediately north of the San Diego River. Other exposures are found to the southeast of the San Diego River around La Mesa and El Cajon. North of Poway Valley and the town of Poway there is a small area of conglomerate about 2 miles long, and to the east ridges of Poway Conglomerate cap several hills south and east of Ramona. The conglomerate undoubtedly also extends south and southeast under a cover of later sediments. The formation is composed chiefly of well-rounded cobbles with sparsely distributed sand lenses and a few bouldery phases. The latter contain boulders ranging up to 10–15 feet and locally appear to be younger channel fillings. The entire section of sediments is exceptionally massive, and structures such as stratification and cross-bedding are extremely vague or lacking. The dip appears to be essentially horizontal, although a low southeasterly dip may prevail in the western portion.

The work on this report was started in the summer of 1951 by Bellemin. His purpose was to make a thorough study of the character of the Poway Conglomerate pebbles. After this preliminary work Merriam joined in the study, which was expanded to include the petrographic relation of the Poway pebbles to the known basement rocks of the area with the intention of determining the origin of the formation. In 1954 a research grant from the Penrose Bequest of The Geological Society of America made it possible to obtain chemical analyses of the Poway pebbles.

Grateful appreciation is expressed to The Geological Society of America for the grant, without which the study could not have been completed, to Dr. A. O. Woodford, whose criticism has been invaluable in this study, and to R. D. Terry, who drafted the chart and geological map.

History of Poway Conglomerate Studies

The first geological mention of the Poway Conglomerate appears in the reports of Blake (1856) on the exploration and survey of a railroad route from the Mississippi River to the Pacific Ocean. During the ensuing period of more than 50 years the rocks were referred to only twice. Both references were made in reports issued by the California Mining Bureau (Hanks, 1886; Fairbanks, 1893). Neither report contributes much information on the nature of these conglomerates.

The first real geologic work done in the San Diego area was a geology and ground-water study by Ellis and Lee (1919). The name Poway was applied to the thick conglomerate beds of the area which were described in general terms but not dated. Two papers by Hanna (1926a; 1926b) on the geology of the La Jolla quadrangle give the stratigraphy of the area and describe the Poway Conglomerate, correlating it tentatively with the Tejon Formation (upper Eocene). Further work by Dusenbury (1932) and Cushman and Dusenbury (1934) on the Foraminifera confirmed this date. Later, a vertebrate fauna was found in the Poway by Stock (1937; 1938); this indicated an Eocene age which agreed in general with that suggested by the Foraminifera. More recently Hertlein and Grant (1939; 1944; 1954) have described the Poway Conglomerate in connection with studies of the San Diego region.

The observations and conclusions of all these authors may be summarized as follows:

(1) Most of the Poway outcrops are north of the San Diego River, between the San Diego River Canyon and Poway Valley; small patches appear in the surrounding areas, particularly to the northeast.

(2) The formation is not of a uniform thickness. The greatest thickness is probably not more than 1000 feet.

(3) The pebbles are large and consist mainly of red porphyry, probably andesite, of unknown source.

(4) The formation contains both marine and continental facies with a probable interfingering of beds very much like parts of the Sespe Formation in Santa Barbara County, California, 200 miles northwest of the Poway area.

(5) It is of Upper Eocene age and equivalent to the Tejon in California, Upper Claiborne of the Gulf Coast, and approximately equivalent to the Bridger Upper Middle Eocene.

Regional Geology and Stratigraphy of the Area

The Poway Conglomerate lies on the west slope of the Peninsular Range, the chain of

mountains extending north and south as the backbone of southernmost California and northern Baja California. The range is a geological unit consisting of the composite Peninsular Range batholith and associated metamorphic rocks. The relative proportions of these two groups differ from place to place, but in general the batholithic rocks greatly predominate.

The metamorphic rocks are complex and of variable composition. A group of metasediments represented by slates, sericite schist, quartz-mica schist, quartzite, and other types is widespread. Metavolcanic rocks termed the Black Mountain Volcanics by Hanna (1926a) and Santiago Peak Volcanics by Larsen (1948) compose another group. They are chiefly mildly metamorphosed flows and pyroclastics of andesite and quartz latite with some rhyolite and basalt. Most of these are dense, fine-grained, gray to black rocks with aphanitic groundmasses. The rhyolitic members are rare and have been described as flows or sills. They are white to greenish white and have abundant phenocrysts of quartz and plagioclase (andesine to oligoclase). Small intrusive bodies of related rocks occur with the volcanic rocks. Most are fine-grained porphyritic granodiorite.

In the Corona and Elsinore quadrangles, about 50 miles north of the Poway outcrops, an area of 20 square miles is underlain by the Temescal Wash quartz latite porphyry. This is a dark-gray to black rock which has a dense aphanitic groundmass and prominent phenocrysts of quartz and andesine. It is considered to be a pre-batholithic rock.

Except for the cover of alluvial patches, small areas of late Tertiary volcanic rocks, and the Poway and related sediments in the west, the batholith is well exposed. Individual intrusions within the batholith range from gabbro to granite. These intrusions form a group showing systematic variations in mineral composition, chemical composition, and sequence of emplacement and are therefore believed to have formed from a primary gabbroic magma by crystal differentiation possibly modified by limited assimilation. Of the known batholith more than half is tonalite, one-third granodiorite, and 2 per cent granite. Extensive unmapped eastern parts are high-silica, low-potassium tonalites and granodiorites.

Texturally all members of the batholith are generally medium-grained granitic, and only a few rocks, principally granodiorites, are fine-grained granitic and nearly aplitic. Granophyric texture occurs in only one small body near Corona (50 miles north of Poway Valley), and aplitic rocks, although widespread, are relatively very minor in volume.

Several parts of the batholith have been repeatedly dated as about 100 million years old, by several element-transformation methods. The geologic age is, therefore, probably Cretaceous (Larsen *et al.*, 1954; Jaffe *et al.*, 1955).

All rocks of the batholith are notable for the unmetamorphosed condition of minerals and textures. Except for minor deuteric changes affecting some of the earliest ferromagnesian minerals and incipient alteration of feldspars in the granodiorites and granites, the rocks are unaffected.

The eastern, topographically higher portions of the Poway Conglomerate lie directly on an irregular bedrock surface, but to the west the conglomerate lies on older marine sediments. The stratigraphy of these sediments has been summarized by Hertlein and Grant (1954). The oldest exposed sediments are Upper Cretaceous, in part at least Campanian gray to brown shales and well-cemented sandstones, which have an approximate thickness of 500 feet. However, deep wells penetrate a conglomerate which may be older, possibly Trabuco (Upper Cretaceous) in age.

Unconformably overlying the Cretaceous is the Eocene La Jolla Formation, composed of three conformable members as follows: lower, Delmar Sand 100 feet; middle, Torrey Sand, 20 to 200 feet; upper, Rose Canyon Shale, 300 feet. Where the Poway rests on the La Jolla Formation the contact is in most places conformable but locally unconformable.

The Poway is locally covered unconformably by more than 1000 feet of soft-gray to yellow marine sands of Pliocene age termed the San Diego Formation. This, in turn, is overlain by the Upper Pliocene to Pleistocene Sweitzer Formation composed of reddish-brown conglomerate with sandy streaks. The Sweitzer is less than 30 feet thick, caps marine terraces in the vicinity of San Diego, and consists largely of reworked Poway pebbles and cobbles.

Pleistocene marine-terrace deposits 20–30 feet thick are common near the coast.

Stratigraphic Column of the San Diego Area

Sweitzer Formation 20 feet	Interbedded reddish-brown gravels and beach sands	Pliocene and Pleistocene
	Unconformity	
San Diego Formation 1000 feet	Coarse and fine-grained yellow to brown sandstones with stringers of gravel	Pliocene

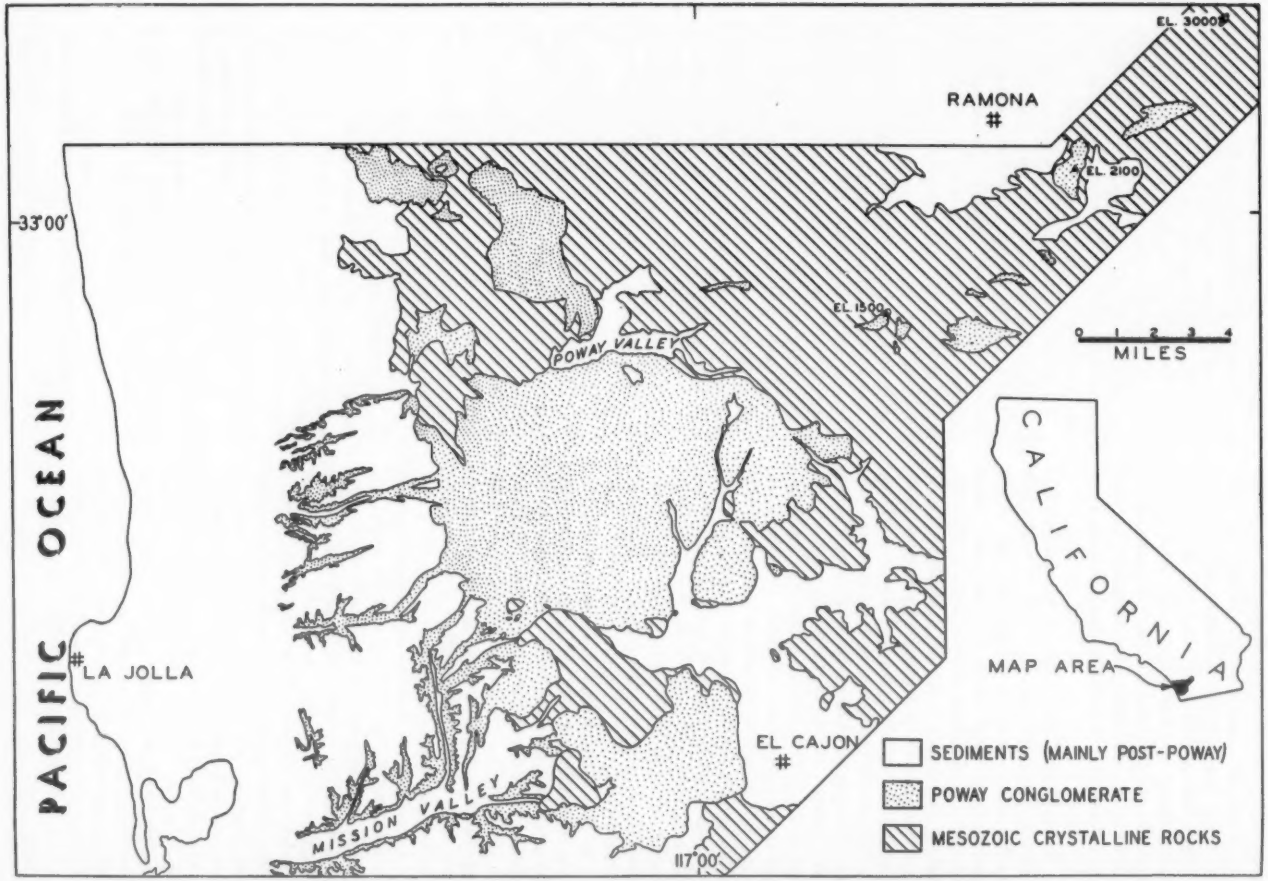

FIGURE 1.—GEOLOGICAL MAP SHOWING DISTRIBUTION OF THE POWAY CONGLOMERATE

	Unconformity	
Poway Conglomerate 1000 feet	Massive conglomerate of well-rounded pebbles and some boulders, consisting largely of rhyolitic tuff and porphyry; interbedded and cross-bedded with sand lenses of coarse sand and fine silty clay; upper zones contain caliche in many places.	
	Unconformity (?)	
Rose Canyon Shale 300 feet	Massive and laminated brownish siltstone with some members of yellow sand and stringers of gravel	Eocene
	Unconformity (?)	
Torrey Sand 20–200 feet	Massive, coarse, arkosic, white sand, cross-bedded in many places	
Del Mar Sand 100 feet	Greenish-gray, pink, purplish-red sand and shale	
	Unconformity	
Chico (?) Formation 1000–2000 feet	Brown sandstones, shales with some bluish-gray mudstones and gray or black fossiliferous shale	Cretaceous
	Unconformity	
Trabuco (?) conglomerates	Reddish-brown sandstones and conglomerates	
	Unconformity	
Batholith of Southern California	A composite, quartz-plutonite batholith consisting of about 20 distinct rock types	
	Intrusive contact	
Black Mountain Volcanics (Santiago Peak Series?) 2000 feet	Extrusive rocks, flows that are primarily andesitic and pyroclastic rocks, metaconglomerates, sandstones, and shales	Jurassic (?)

FIGURE 1.—GEOLOGICAL MAP SHOWING DISTRIBUTION OF THE POWAY CONGLOMERATE

POWAY CONGLOMERATE COLLECTIONS

Method of Study

The samples for pebble counts were obtained in about the same way at each locality. The usual sample consisted of a complete sample of the conglomerate, matrix and pebbles, with not less than 100 pebbles. Fifteen such samples of the Poway Conglomerate were collected; of these only 3 had less than 100 pebbles, and 4 had more than 200 pebbles.

The sample localities (Fig. 2) were selected as typical of the local section. A small hole (10 by 10 by 8 inches) was then dug, and all the material from the hole was put into a sack or box. Where the conglomerate was very well cemented the collection was made from a larger surface area by picking at the surface and collecting all the pebbles knocked loose. Before leaving a locality a check was made in the general area to see if any significant features or clasts of unusual composition were apparent in the outcrop. At no time were random samples collected for a pebble count.

The sample was soaked in water to loosen the pebbles from the matrix; then it was passed through a screen with a quarter-inch mesh. The material that passed through the screen was considered as the matrix. The longest diameters of all clasts on the screen were then measured with a steel ruler or 6-inch calipers. The small pebbles, with lengths between 1 inch and a quarter of an inch, were checked for significant features but otherwise not included in any further study.

All the pebbles more than 1 inch long were broken and studied with a hand lens. The hand-lens determinations were adequate to distinguish most of the rock species. However, in the rhyolitic rocks albite was rarely distinguishable, primarily because the very fine discontinuous twinning was obscured in most rocks by alteration of the feldspar. From this study the rock species were determined, and typical specimens for thin sectioning were selected.

The Poway Conglomerate collections described are numbered by collection localities in the sequence in which they were taken and recorded. They are numbered PC (Poway Conglomerate) 1 to 19. At locality PC-2 no collection was made, and at PC-10, 11, and 12 there was some question of whether the material was Quaternary alluvium or Poway, and therefore these four localities were not included in the study. The collecting localities are shown on the accompanying map. Table 1 summarizes the pebble count for each locality.

Description of Poway Conglomerate Collections

POWAY CONGLOMERATE SAMPLE PC-1: The first sample was taken on the south side of the San Diego River in an area mapped as Poway Conglomerate by Hanna (1926a). In this area Poway Conglomerate is present on both sides of the San Diego River. The sample was collected in a gully 20 feet off the west side of Fairmount Road 1 mile north of El Cajon Boulevard. One long stringer of sand shows that the beds are practically horizontal.

The size distribution of the clasts indicates a maximum at 1 or 2 inches in diameter. The largest boulder is about a foot long. Nearly all pebbles and larger clasts are well rounded and ellipsoidal in shape.

The matrix consists of angular grains, ranging in size from grit to fine sand. Quartz and feldspar are the most abundant minerals, about

equally divided, and amounting to 95 per cent of the matrix. Most of the remaining 5 per cent is weathered biotite. Most of the quartz grains are clear, and many have a roughly hexagonal shape, which suggests that their probable source was a phenocryst weathered free from a porphyry.

1½ inches in diameter and a larger piece of a tooth resembling a molar imbedded in bone. Both of these were in a sandy layer and were weathered to the point where they crumbled to the touch.

POWAY CONGLOMERATE COLLECTION PC-3: Collection PC-3 was obtained beneath a Pleisto-

TABLE 1.—SUMMARY OF POWAY CONGLOMERATE PEBBLE COUNT

	PC-1	PC-3	PC-4	PC-5 & 9	PC-6	PC-7	PC-8	PC-13	PC-14	PC-15	PC-16	PC-17	PC-18	PC-19
Rhyolitic porphyry or tuff	156	101	103	168	92	102	193	178	43	56	88	86	130	170
Felsitic rocks	11	9	77	0	4	8	8	4	15	8	5	6	10	15
Aplitic rocks	7	7	1	1	2	3	3	12	0	1	4	1	1	9
Granitic rocks	30	0	0	16	6	0	3	8	1	1	6	0	3	7
Quartzite	12	24	8	2	2	44	15	3	1	0	4	7	2	2
Andesite porphyry	10	0	4	8	0	0	10	2	0	0	3	0	0	0
Miscellaneous and unknown	10	12	20	37	14	24	39	24	3	6	1	9	7	29
Total	236	153	153	232	120	181	271	231	63	72	111	109	153	232

The pebbles are 66 per cent rhyolitic rock; most other samples are even higher in this constituent. The most interesting specimens of the granitic group were six pebbles that had a roughly concentric structure and uneven texture. These may be parts of orbules or inclusions. Unfortunately all the orbicular specimens were weathered and crumbled away. Most of quartzite is light brown. The felsitic material is light and dark, and the darker material resembles basalt. The aplitic pebbles were classed as such on the basis of the sugary texture. Plagioclase is abundant enough in some to permit classifying them as granodiorite aplites.

Table 2 gives a pebble count on small material from the same locality. Even though the texture could not be determined in each specimen, the porphyry (or welded tuff) is dominant, but it has diminished with respect to the other rock types. The greater abundance of siliceous "porphyries" among the larger clasts is perhaps due to the great resistance of the rhyolite group compared to a more easily fractured rock which might shatter and produce many small fragments, while the rhyolite remains whole. This also shows that the larger clasts, more than 1 inch, are reliable indicators of the character of of the whole conglomerate.

About 75 feet up canyon on the same side of the gully some vertebrate fossil material was found. It consisted of a tusklike tooth about

cene terrace on the north side of Carroll Canyon north of Miramar, on the former course of U. S. Highway 395. The vertical exposure of conglomerate is well cemented; blocks of conglomerate that break away, because of undercutting, hold together in large blocks. The thin red adobe Pleistocene soil penetrates into the conglomerate and colors the top 2 or 3 feet. The beds are apparently roughly horizontal.

The size distribution of the pebbles is approximately the same as at PC-1, but the matrix is different. There is an abrupt break in the sorting of grains less than a quarter of an inch in size with a silty-sandy clay as the matrix.

The rhyolite porphyry and tuff described in PC-1 is again found here in almost the same percentage of the total. The felsitic rocks are about as abundant as in PC-1. One specimen shows flow structure. The aplites are about the same as in the previous collection. A specimen that resembles a lamprophyre (dark gray) was included in this group.

POWAY CONGLOMERATE COLLECTION PC-4: Sample PC-4 was collected from the hill at the top of the Poway grade on the former course of U. S. Highway 395. The hilltop (elevation 1022 feet) on the east side of the road is part of the ridge that is cut by the road. The conglomerate is well cemented in a sandy clay matrix. The bedding is roughly horizontal, based on some sand lenses visible in the road cut.

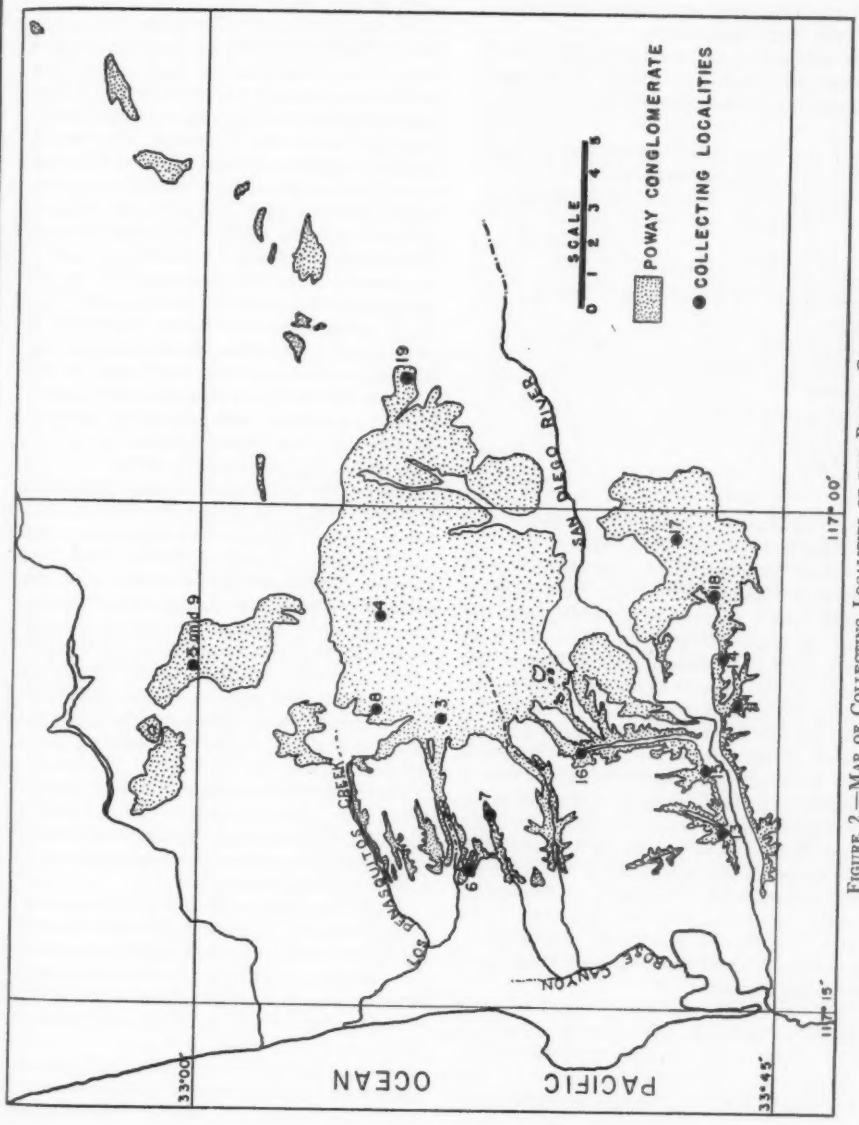

FIGURE 2.—MAP OF COLLECTING LOCALITIES OF THE POWAY CONGLOMERATE

The size distribution of the pebbles indicates a tendency toward the smaller sizes. Pebbles 1 inch and 1¼ inches in diameter amount to 55 per cent of the total sample.

In addition to the usual rhyolitic material, several rarer forms were found. These include four specimens that show distinct flow structure, two that may be granite porphyry, and one that looks like an agglomerate.

TABLE 2.—TEXTURE CLASSIFICATION OF SMALL PEBBLES FROM PC-1
(A quarter of an inch to 1 inch in diameter)

	Specimens	Per cent
Porphyritic (or welded tuffaceous) texture	111	49
Felsitic texture	35	16
Granitic and schistose texture	34	15
Quartzite	16	7
Miscellaneous and unknown	29	13
Total	225	100

POWAY CONGLOMERATE COLLECTIONS PC-5 AND PC-9: Collections PC-5 and PC-9 were obtained from a road cut of U. S. Highway 395 south of Lake Hodges and about 4.8 miles north of the Los Penasquito Creek bridge. This is one of the northernmost exposures of the Poway Conglomerate in this north-south section. At this point unconformities within the Poway Conglomerate are visible. Truncated sand lenses and gravel beds are clearly seen in the road cut. One large mass of greenish sandy silt material similar to the greenish layers in the Sespe Formation of the Santa Ana mountain area occurs in the conglomerate. The beds are roughly horizontal. Two samples were collected.

PC-5 was collected on the ridge top on the west side of the road cut. PC-9 was collected on the same side of the road in the cut, about 3 feet above the road level.

The size distribution repeats the distribution pattern of previous collections; more than a third of the sample consists of 1-inch pebbles. The matrix is a medium to fine sand with some clay. Most of the sand grains are subangular and consist of quartz, feldspar, epidote, ferromagnesians that are weathered, and bits of groundmass from the porphyries.

This is the only collection in which no felsitic clasts were found. The aplites were represented by only one specimen, and the quartzite by only two.

POWAY CONGLOMERATE COLLECTION PC-6: PC-6 was collected about 100 yards to the north of the Sorrento-Linda Vista Road 2.7 miles from Sorrento. (There are two Linda Vistas within 6 miles. This refers to the northern Linda Vista, which is outside the San Diego city limits.) The ridges on both sides of the road are capped with Poway Conglomerate which overlies yellow sands very irregularly. The contact is almost vertical for about 2½ feet. The attitude of the conglomerate is roughly horizontal.

The size distribution of this sample indicates a decrease in the number of 1-inch pebbles, though they still amount to about one-third of the sample. The general appearance of this conglomerate gave the impression that it was better sorted than the others. However, in general, it does show roughly the same size distribution. The largest boulders seen at this point were about 2 feet long, but there seemed to be no boulders between 7 or 8 inches in diameter and the 18-inch to 2-foot size. This peculiarity was noted again at PC-8.

POWAY CONGLOMERATE COLLECTION PC-7: Collection PC-7 was made on the mesa between Linda Vista and Miramar, about 150 yards off the south side of the dirt road from Sorrento (Sorrento-Miramar road), near the intersection with the Linda Vista-Miramar road, adjacent to the north boundary of Kearny Field. The surface of the mesa is covered with an adobe clay soil. Gullies of various sizes dissect the mesa; one of these small gullies (6–8 feet deep at this point) provided the exposure that was sampled. No bedding is visible in the area.

The size distribution shows a maximum at the 1-inch size which amounts to 30 per cent of the total. No very large boulders were visible in the gully; the largest were approximately a foot at the longest diameter.

In general, the variety of rock species is not so large as usual. More than half of the total specimens collected were rhyolitic material. The quartzite, as indicated on the surface of the outcrop, is very abundant. This is the largest amount, 24 per cent, found in any of the collections.

On the near-by mesa surface well-polished black pebbles, which resemble basalt, were evident but not common. They are noticeable because of the black contrast with the red soil and the fact that black rocks are scarce in these collections. Half a dozen of these were collected at random over an area 100 yards in diameter. The specimens are sparsely porphyritic and have a black crystalline, almost aphanitic,

groundmass. The "phenocrysts" are clear quartz, and the groundmass under the microscope proved to be mostly black tourmaline. This rock may be called a quartz tourmalinite. It was not found in the 181 pebbles of the PC-7 collection.

POWAY CONGLOMERATE COLLECTION PC-8: One of the largest samples collected is sample PC-8, which has 271 pebbles and cobbles. This sample was obtained on a road cut of U. S. Highway 395, 1.8 miles north of Carroll Canyon. In this area the surface is gravel-covered, at the edge of rolling hills which eventually rise to Black Mountain (1573 feet) and Twin Peak (1312 feet), the highest peaks in the area.

In the road cuts and gullies the Poway Conglomerate shows numerous sand layers. Where the collection was made the interbedded sands and gravels show much cross-bedding, none of which was clear enough to determine direction of origin. The sands and gravels appear to be well sorted and form definite layers. The beds dip so slightly that it is difficult to obtain a definite attitude. The average dip is between 5° and 0°, and the strike is between 5° and 15° east of north.

The size distribution shows a heavier concentration of pebbles around the 2-inch size than in any previous collection. Also, 31 clasts were more than 3 inches long. The largest was 5¼ inches long. Along the 400 feet of road cut the three largest boulders were: 19 inches (quartzite), 24 inches (quartzite), and 24 inches (porphyry). These were all flat ellipsoids with smoothly polished surfaces. The quartzites are covered with bulb of percussion fractures, which were produced by being struck violently by other boulders.

The matrix of this sample is sand. The grain size ranges from coarse to fine with very little clay. The predominant mineral is quartz, and feldspar is moderately abundant. Certain layers of the sandy section of the road cut were rich in dark minerals, which have produced a dark-gray sand.

The most abundant rocks are, again, the rhyolitic group. They include one possible quartz latite, a tourmaline quartz porphyry, one agglomerate (?), and granitic porphyry which amounted to 16 per cent of the rhyolitic group. Quartzite amounted to 6 per cent of the total and occurred in pink, white, and brown varieties. The miscellaneous and unknown group contains eight types that were represented by only one or two specimens each, such as a gneiss (?), two different porphyries, and various doubtful types.

POWAY CONGLOMERATE COLLECTION PC-13: Collection PC-13 was made in a gravel quarry along the west edge of the old road from Friars on the north side of the San Diego River to Linda Vista (within the San Diego city limits) on the top of the mesa. The whole area consists of Poway Conglomerate, which is exposed in all the canyons, gullies, and the gravel quarries. Many lenses of cross-bedded sands are visible in the cuts. The conglomerate is moderately well cemented with sand and clay. The discontinuous sand layers indicate a practically horizontal attitude.

This rather large collection of 231 pebbles has 29 per cent of 1-inch pebbles, about the same as usual.

The matrix is a sand of subangular grains; it consists largely of quartz, feldspar, epidote, dark undetermined minerals, rock fragments, and some silt.

POWAY CONGLOMERATE COLLECTION PC-14: Collection PC-14 consists of only 63 pebbles collected on the vertical road cuts on the north side of Mission Valley road about 600 feet east of the intersection with Fairmount Avenue.

The only unusual feature of this collection is the rarity of the reddish groundmass in the porphyry. The other differences are, perhaps, due to the small sample. The black quartz-tourmaline rock in the miscellaneous group is the same as the black rocks found at PC-7 and PC-13.

POWAY CONGLOMERATE COLLECTION PC-15: Collection PC-15 is also rather small and could be considered part of the PC-13 collection. Both collections are from the same zone of beds along the north side of the San Diego River, about 3 miles apart. The PC-15 collection was made 1 mile west of Murphy Canyon at the base of a quarry surface. The top zone is rich in caliche at this point. The matrix in this area is a coarse sand.

The striking feature of this collection is the lack of quartzite and, less significantly, of andesite, and the small amount (one specimen each) of granite and aplite. Part of this irregularity may be explained by the small sample As the major constituent, the rhyolitic rocks make up the usual high percentage. The somewhat unusual proportions of the minor constituents probably indicate only that the sample is too small to make these proportions significant.

POWAY CONGLOMERATE COLLECTION PC-16: Collection PC-16 was made in Murphy Canyon near the top of the mesa in a road cut 4 miles north of Friars Road. The Poway Conglomerate is exposed on both sides of the canyon

from the San Diego River to the top of the mesa. Numerous layers of sand and silt occur in the conglomerate.

About 500 feet south of the collecting point some marine fossils (pelecypods) occur in a silty zone. Unfortunately they were so weathered that they crumbled to the touch.

The matrix is a coarse to fine sand with no clay. Clear quartz is abundant, and feldspar and rock fragments are also present.

In this collection the rhyolitic group amounts to almost 80 per cent of the sample. The felsites and aplites both show a light and dark facies. The pink orthoclase is particularly striking in the aplites.

POWAY CONGLOMERATE COLLECTION PC-17: Collection PC-17 has only 109 pebbles, without matrix. It was obtained from a ridge top northwest of Lake Murray Boulevard about 2.2 miles from El Cajon Boulevard. Most of the ridges in this area are capped with conglomerate, and in most places it is weathered (in part) to a red soil.

This collection lacks granite. Andesite is also missing. The quartzites are black quartz (schist) and a red and black quartzite.

POWAY CONGLOMERATE COLLECTION PC-18: Collection PC-18 was made on Mission Road in Alvarado Canyon near the top of the mesa just east of the igneous material. Some of the beds in this area may be Poway Conglomerate with a cap of Poway Conglomerate alluvium. The road cuts here are largely in the soil horizon of the Poway Conglomerate, and most of the sandy matrix is soil. Some of the boulders and cobbles show two stages of weathering.

This collection has the largest percentage (85 per cent) of rhyolitic material of all the collections. Two pebbles in the group show flow structure.

The matrix is poorly sorted, and grains range in size from coarse sand to clay. Quartz predominates, and much of it is clear. Rock fragments of sand size are common. The clay has a greasy feel and molds like modeling clay.

POWAY CONGLOMERATE COLLECTION PC-19: Collection PC-19 was made near the western margin of the main body of the Poway Conglomerate. It was collected from a road cut on the east side of the road from Lakeside to Ramona, 3.1 miles north of the Barona turnoff. This is near the contact of the Poway Conglomerate and the local Mesozoic crystalline rocks.

The matrix is predominately angular, coarse sand but also includes finer sand and silt and very little clay. The matrix amounts to less than a quarter of the total by volume. The composition of the sand is roughly 50 per cent quartz, 45 per cent weathered feldspar, 5 per cent weathered biotite, and other dark minerals.

This sample is about average. Andesite is absent. One specimen of black quartz-tourmaline rock was found.

Summary of All the Poway Conglomerate Collections

An average of the Poway Conglomerate characteristics can be determined by considering all the samples as part of one composite sample. The total number of pebbles classified in the collections just discussed is 2317. This provides a large sample in which nearly all areas of the conglomerate are represented.

A size-distribution average would probably not be valid if a definite geographic gradation trend could be seen; however, none is discernible.

About 9 per cent of the conglomerate pebbles are 3 inches or more in diameter, 19 per cent are 2–3 inches, and 72 per cent are 1–2 inches, if one may judge from 2000 pebbles measured with steel scale or calipers, from the various localities reported (Table 3). The mean size is between 1 and 2 inches maximum diameter. Most publications give larger mean sizes (Hanna, 1926a).

The largest observed boulders were about 2 feet in diameter. These, and slightly smaller ones, occurred at many localities. Usually they were well-rounded, ellipsoidal boulders of quartzite or rhyolite porphyry. Larger boulders have been reported by others, but none was seen.

The matrix of the conglomerate ranges from a silty clay to a coarse sand. In the pure conglomerate beds it probably amounts to a quarter or a fifth of the total by volume. In a few places a zone of tightly packed pebbles occurs with very little matrix between the pebbles.

The clay matrix resists blows of the pick because it is not brittle and in many cases will not soften in water. The clay is usually mixed with silt or mica flakes.

The sandy matrix is poorly sorted and has grains ranging from grit to silt size; medium or coarse sand usually predominates. Most of the mineral constituents are quartz and feldspar and have 5–15 per cent ferro-magnesian minerals and rock fragments. The quartz is nearly always clear and in places shows remnants of the hexagonal crystal form. Most of the feld-

spars are weathered, but in a few a fresh cleavage chip appears. The minor constituents are biotite, tourmaline (?), epidote, magnetite, and many grains of porphyry groundmass.

Table 4 shows the overwhelming abundance of the rhyolitic material which ranged from a

TABLE 3.—AVERAGE SIZE OF THE POWAY CONGLOMERATE PEBBLES

Size	Total pebbles from 12 localities	Per cent
3-inch maximum diameter or larger	179	8.95
2- to 3-inch diameter	373	18.65
1- to 2-inch diameter	1448	72.40
Total	2000	100.00

low of 56 per cent at PC-7 to a high of 85 per cent at PC-18. This rock occurs in several varieties of color and texture. The most abundant variety, which makes up 32 per cent of the 1666 specimens, is brown (light to dark brown). The most striking variety, amounting to 25 per cent of the total, is red or red brown. About half of these specimens contain piedmontite, and the other half contain iron oxide; in some cases both minerals are present in the same specimens. About 22 per cent of the rhyolitic porphyries have a gray groundmass. Most of the remaining 21 per cent have mottled or discolored groundmasses because of weathering. The textural varieties identifiable wih the hand lens differ largely in the matter of the abundance of phenocrysts. In some cases they are sparse, 5–10 per cent of the rock, and in some cases they make up about 80 per cent of the rock. Flow structures are visible on a few specimens.

The next most significant rocks are the felsites, which amount to about 5 per cent of the color variety as the rhyolitic types and are probably related.

The quartzite abundance ranges from 24 per cent at PC-7 to zero at PC-5 and PC-15. These also occur in several color varieties such as white, gray, pink, and brown. The variety probably indicates different sources.

The remainder of the rocks occur in such small quantities that it is difficult to make any general observation concerning them. The aplitic, granitic, and andesitic rocks vary in number and character throughout the Poway Conglomerate. The lack of homogeneous character of this group makes it difficult to do more than give the percentages of the groups rather than specific information concerning one species of rock. A very fine-grained light-gray granite that may be slightly metamorphosed occurs in

TABLE 4.—AVERAGE PER CENT OF ROCK TYPES IN THE POWAY CONGLOMERATE
(Based on all the localities sampled)

Rock species	Number of specimens	Per cent
Rhyolitic porphyry and tuff	1666	72
Felsitic rocks	120	5
Aplitic rocks	52	2
Granitic rocks	75	3
Andesitic rocks	37	2
Quartzite	126	6
Miscellaneous and unknown	241	10
Total	2317	100

several collections. It amounts to nine specimens or 6 per cent of collection PC-9, three specimens or 1 per cent of collection PC-13, and one specimen or .5 per cent of collections PC-18, and PC-19. A distinctive rock placed in the miscellaneous group is the black quartz-tourmaline rock. One pebble was found in the collection at each of the following localities: PC-13, PC-14, PC-17, and PC-19; it was visible on the ground at PC-7 but was not found in the sample. Most of the 10 per cent of the miscellaneous and unknown group are weathered, and recognition is difficult or impossible.

PREDOMINANT ROCK TYPES OF THE POWAY CONGLOMERATE

Sodic-rhyolitic Poway Suite

In general, most of the Poway clasts came from a distinct group of volcanic rocks, all of which are characterized by (1) high silica content, (2) sodic plagioclase, (3) altered ferromagnesians, (4) mild alteration and recrystallization, (5) pyroclastic and porphyritic textures closely resembling one another. Many of the nonvolcanic rocks also have one or more of these properties.

The most notable mineralogical feature of the rhyolites is their plagioclase, which generally ranges from An_2 to An_{10}. This mineral, which invariably shows incipient alteration to

sericite, is everywhere unzoned. The ferromagnesians, largely biotite but largely hornblende, are almost completely altered to iron-ore pseudomorphs with or without sericite. Some of the rocks have been slightly tourmalinized, and specimens showing secondary piedmontite, although not common, are widespread. Highly corroded phenocrysts and crystal fragments of quartz are typical. Potassium feldspars are uncommon, but a few rocks were found containing 5 or 10 per cent orthoclase or perthite. The groundmasses and fine-grained lithic fragments appear to have been recrystallized to a fine equigranular mass probably composed of quartz and feldspar.

Textures of the rhyolitic rocks are clastic and nonclastic. The similarity of these two types plus slight recrystallization makes it difficult, and in many cases impossible, to determine to which group a rock belongs. Those believed to be pyroclastic are characterized by the following features: (1) prevalence of lithic fragments and broken crystals, (2) relics of fluxion structure typical in welded tuffs, (3) wide range in size of crystals and crystal fragments.

Porphyritic and seriate porphyritic textures occur but are less common than clastic textures. Rocks of the former group lack lithic fragments and broken crystals and have 1- to 2-mm phenocrysts enclosed in a uniform microgranular groundmass.

All rhyolitic rocks have been classified as soda rhyolite porphyry or soda rhyolite tuff. Some are indeterminate, but for the solution of the problem at hand exact identification is not essential. More important is the question of whether or not their chemical composition has been changed by silicification or albitization. Slight recrystallization, incipient sericitization, and lack of zoned palgioclase are indicators of such changes, which may have been primarily deuteric. Evidence against silicification lies in the close correspondence between composition of rhyolite and granophyre; the latter shows no signs of silicification. The chemical analyses indicate probable silicification of one rock, PC-19 (Table 5, analysis 8). The points plotted for the oxides of this rock on Figure 3 do not conform to the curves produced by the other analyses. Since this rock is a soda rhyolite tuff, it is reasonable to assume that it would have sufficient pore space to permit addition of 10 per cent silica. Recalculations on this basis result in oxide points that fit fairly well.

Granophyres

Granophyres occur in the coarse portions of the typical Poway but are much more prevalent in the patches of coarse, young gravels locally overlying the typical phase in the east. In the latter it is present as large subangular to subhedral boulders ranging up to several feet in diameter. A slightly weathered, discolored zone an inch thick is common, but most of the rock is fresh and hard.

Most specimens tend to be uneven-textured and somewhat porphyritic, although others are entirely micrographic intergrowths. Phenocrysts averaging 2–3 mm are prominent, possibly because of weathering. Interstitial material is micrograngular to aplitic or micrographic or, more commonly, a combination of these.

Quartz is one of the principal minerals; it generally exceeds 50 per cent as phenocrysts, intergrowths, or interstitial granules.

Microperthite is the dominant feldspar. It typically contains not more than 10 per cent albite and shows incipient alteration.

In addition to its occurrence in perthitic intergrowth albite is usually present as individual rectangular crystals, which are in most cases phenocrysts 1–2 mm long. The mineral is everywhere unzoned and charged with minute inclusions, which are probably the result of mild alteration. The composition ranges from An_2 to An_{10}.

Ferromagnesian minerals are very sparse and highly altered, principally to iron ore. Where the original mineral can be recognized it is usually biotite.

Minor accessories other than iron ore are very rare.

Aplites

A substantial percentage of the younger, coarse gravels consists of rocks in which the hand specimens resemble aplites. The textures are somewhat uneven xenomorphic granular and have gradations to sericite porphyritic, granophyric, or fine granitic. The largest grains in the aplites rarely exceed 2 mm, and most are smaller.

Anhedral to subhedral equidimensional grains of microperthite compose nearly half of the average aplite. It is low in albite and slightly altered. Quartz anhedrons of various sizes amount to approximately 45 per cent. Small individual crystals of plagiolase rang-

ing from An_5 to An_{10} are present in small amounts. Biotite is the only dark mineral, except for iron ores, and is very sparse, considerably altered to opaque grains, and in small flakes. Varying degrees of freshness are found, but nearly all minerals show abundant fine inclusions.

Andesitic Types

These represent a rather uncertain group. The rocks included are nearly all brown porphyritic-aphanitic rocks, which have lathlike feldspar that in some cases cannot be determined because of alteration. On the basis of a few hand-lens determinations several rocks were grouped together. They occur in only about half of the localities studied and amount to less than 2 per cent of the total.

Quartzites

The quartzites show a great variation in variety and abundance. The abundance ranges from 24 per cent in one sample to none in two samples. In texture, most show the typical quartzite structure of even granular quartz grains, but many have schistose structure. The colors range from pure white through cream, tan, pink, brown, to very dark gray, almost black. The dark specimens have a more schistose texture.

Chemical Compositions of the Poway Rhyolitic Rocks

Chemical Characteristics of the Poway Rhyolitic Suite and Its Relation to the Peninsular Range Batholith

Eight of the most abundant and typical rocks of the conglomerate were analyzed. (Tables 5, 6). The writers believe that at least 70 per cent of the cobbles of the formation are identical to, or very closely resemble, these eight. Although some of the analyzed rocks are pyroclastic they probably formed rapidly as welded tuffs without appreciable contamination and thus represent the composition of the magma from which they were derived.

A comparison of the analyses with those of batholith rocks was made to determine similarities or differences indicative of genetic relationships. Such a comparison is most readily seen on a plot such as Figure 3, in which curves for the plutonic rocks of the Peninsular Ranges (Larsen, 1948) have been extended to accommodate the Poway rocks. In all cases the extensions appear to be reasonably conformable with the remainder of the curve and follow trends to be expected from normal magmatic differentiation. The values for FeO and K_2O do not fit so well as those for other components. The deviation of the former is caused by oxidation of iron by weathering and shallow intrusion or extrusion. Possibly the configuration of the K_2O curve for the batholith rocks is not valid, since there is a wide spread of points at this end. A smoother curve would pass through the middle of the K_2O points for the Poway rocks.

In plots of Poway rocks, made on curves of Sierra Nevada and other batholiths, the deviations were considerably more than in the case described above.

Trace-element studies have been made of the batholith rocks, but the trace-element content of the Poway rocks is unknown, with the exception of SrO and BaO, which were determined for rocks PC-7 and PC-8. PC-7 yielded 700 ppm of BaO and 100 ppm SrO; PC-8 yielded 2200 ppm BaO and 170 ppm SrO. These figures are in good agreement with those of batholith rocks (Larsen, 1954), p. 139) and are distinctive.

The age of the batholith has been determined by at least two methods based on radioactive disintegration. However, such methods are not applicable to clasts of the conglomerate, for weathering may have modified their chemistry. Furthermore, accessory minerals such as zircon are extremely rare.

Petrography of Analyzed Rocks

PCM-1, GRANITE APLITE: In hand specimen the rock is massive, white to light tan. The prominence of 1- to 2-mm quartz grains produces a porphyritic appearance, although it is more evenly grained.

Under the microscope the texture is fine-grained hypautomorphic to xenomorphic granular; some portions are porphyritic and others granophyric.

Microperthite in anhedral and subhedral grains and in graphic combinations with quartz composes 50 per cent of the rock. Quartz as anhedrons and in intergrowths amounts to 45 per cent. In addition to the albite of the perthite scattered small individual grains of plagioclase (An_8) are present.

Ragged flakes and groups of flakes of biotite occur (2 per cent) with associated grains of iron ore. Smaller grains of the latter mineral are widespread. All minerals appear fresh, although dustlike inclusions are found throughout.

PCM-6, GRANOPHYRE: The hand specimen is light brown to white, massive, and appears porphyritic because of abundant, poorly distributed quartz and feldspar crystals.

Biotite, always accompanied by iron ore, occurs as flakes and groups of flakes which have extreme pleochroism (X′ light yellow, Z′ nearly black). Roughly 2 per cent of the rock is biotite,

TABLE 5.—ANALYSES OF TYPICAL POWAY ROCKS

	1	2	3	4	5	6	7	8
SiO_2	77.26	77.41	76.09	76.55	74.75	69.88	76.96	79.75
Al_2O_3	12.44	12.49	12.45	12.59	12.75	15.12	12.11	10.78
Fe_2O_3	.90	.73	1.50	.97	2.29	3.23	1.43	1.36
FeO	.10	.10	.21	.21	.20	.18	.17	.14
MgO	.06	.06	.19	.20	.12	.66	.11	.19
CaO	.20	.34	.08	.06	.20	1.34	.19	.21
Na_2O	3.70	3.44	3.59	1.96	2.80	4.38	3.32	3.07
K_2O	4.39	4.41	4.55	5.64	5.26	3.48	5.12	3.65
H_2O+	.26	.33	.39	.80	.77	.60	.31	.47
H_2O-	.12	.17	.08	.18	.15	.09	.10	.13
CO_2	.01	.03	.02	.03	.01	.01	.03	.01
TiO_2	.08	.09	.15	.16	.17	.38	.15	.14
P_2O_5	.02	.01	.03	.03	.02	.10	.02	.00
MnO	.01	.01	.02	.01	.01	.04	.01	.03
S	.02	.03	.02	.02	.02	.04	n.d.	n.d.
Cl	n.d.	n.d.	.03	.04	n.d.	n.d.	n.d.	n.d.
SrO	n.d.	n.d.	n.d.	n.d.	.01	.02	n.d.	n.d.
BaO	n.d.	n.d.	n.d.	n.d.	.07	.22	n.d.	n.d.
	99.57	99.65	99.40	99.45	99.60	99.77	100.03	99.93

1. Granite aplite (PCM-1) from coarse phase of Poway Conglomerate in road cut a quarter of a mile south of Fernbrook.

2. Granophyre (PCM-6), same locality as 1.

3. Soda Rhyolite tuff (PC-1) from gully 20 feet off west side of Faimount Road, 1 mile north of El Cajon Boulevard.

4. Rhyolite tuff (PC-9I), road cut on U. S. Highway 395, 4.8 miles north of Los Penasquitos Creek Bridge.

5. Soda rhyolite porphyry (or tuff) (PC-7), 150 feet south of intersection of Sorrento-Miramar road and Linda Vista-Miramar road.

6. Soda rhyolite porphyry (or tuff) (PC-8), road cut on U. S. Highway 395, 1.8 miles north of Carroll Canyon.

7. Soda rhyolite tuff (PC-17), ridge top northwest of Lake Murray Boulevard 2.2 miles from El Cajon Boulevard.

8. Soda rhyolite tuff (PC-19), road cut on Lakeside-Ramona road 3.1 miles north of Barona turnoff.

Under the microscope 80 per cent of the rock is seen to consist of a granophyric intergrowth of quartz and perthite surrounding and in some places radiating from euhedral to subhedral quartz and microcline microperthite crystals.

Quartz individuals are 1- to 2-mm. in diameter and exhibit strained extinction. Microcline microperthite crystals of similar size are filled with inclusions resulting from incipient alteration. Individual plagioclase crystals (An_{10}) are distributed throughout and range from 0.1 to 0.5 mm.

whereas muscovite is much rarer. A few small zircons are present.

Because of intricate textural intergrowths the relative amounts of the chief minerals are difficult to estimate, but quartz is roughly 50 per cent and microperthite 40 per cent.

PC-9, I RHYOLITE TUFF: The hand specimen is a pinkish-gray porphyritic rock with a dense, hard, aphanitic groundmass. Pink potassium feldspar and quartz phenocrysts make up 30 per cent. Ferromagnesian minerals are not visible with the hand lens.

A pyroclastic texture is seen in thin section with a variable microgranular groundmass of grain ranging from 0.05 to 0.5 mm. Irregular-patches and streaks of coarse and finer ground mass material are common. Flow structure appears around some of the larger crystals.

Quartz (50 per cent of rock) is the commonest mineral. The crystals, which are corroded and in a few cases fragmental, commonly show strained extinction.

Microperthite is second in abundance and is a principal constituent in the finer matrix. It is fresh to slightly altered.

Plagioclase is too highly saussuritized for more accurate identification, although it is presumably albite.

Pseudomorphs of iron ore, sericite, and piedmontite replacing biotite are typical but sparse.

PC-7E, SODA RHYOLITE TUFF (OR PORPHYRY): This is a light-brown rock with an outer shell bleached by weathering. Nearly 40 per cent of the rock is phenocrysts which are prominent and reach 6 mm in diameter. The groundmass is massive, dense, and aphanitic. Small rods and aggregates of crystals of black tourmaline are widespread.

Under the microscope the texture is indeterminate, although it is probably pyroclastic rather than porphyritic. The groundmass is microgranular and exhibits wider variations in grain size.

Approximately 50 per cent consists of quartz, which forms corroded phenocrysts with sharp extinction and many fractures, as well as abundant granules in the groundmass. Microperthite phenocrysts (10 per cent) carrying very little albite are fairly fresh but contain numerous inclusions, possibly because of incipient alteration. Unzoned albite, moderately saussuritized and highly fractured, constitutes 10 per cent of the rock. Iron ore is spread throughout.

Tourmaline, apparently secondary, forms rods, rosettes, and irregular patches. It is strongly pleochroic; ϵ is pale pink, and ω is dark blue. Basal sections reveal zoning wherein cores of light blue are surrounded by dark blue.

PC-17, K SODA RHYOLITE TUFF: This rock is light brown to buff on weathered surface but gray where fresh. Phenocrysts are inconspicuous except where weathered. The groundmass is fine, dense, and cut by many cracks.

Under the microscope the texture is pyroclastic and has crystals (35 per cent) ranging up to 1 mm in diameter set in a microgranular matrix. The latter component composes 65 per cent of the total rock and contains local areas approaching granophyric character. Sheared zones, prominent in plane light but invisible in crossed nicols, cut and offset phenocrysts and are largely healed by recrystallization.

Approximately 50 per cent of the larger crystals are quartz; they range from subhedral to anhedral and show strained extinction. Quartz also composes at least half of the fine matrix.

Albite (approximately An_8) forms unzoned, twinned, and slightly sericitized crystals averaging 0.5 mm long. It is present as 10 per cent of the larger crystals but was not recognized in the groundmass. The microperthite contains very little albite, is slightly altered to sericite, and occurs as euhedral to subhedral crystals. It is prominent among fine constituents. Iron ore is widespread, but other accessories are lacking.

PC-8T, SODA RHYOLITE PORPHYRY (OR TUFF): The hand specimen is purplish brown specked with white and gray phenocrysts, which compose nearly half of the rock.

TABLE 6.—NORMS OF ANALYZED ROCKS*

	1	2	3	4	5	6	7	8
Q	39.30	40.00	38.00	44.70	37.75	26.62	37.10	47.52
or	26.13	26.13	26.13	33.36	31.01	20.59	30.16	21.61
ab	31.44	28.82	30.29	16.77	23.70	37.00	28.05	25.94
an	1.11	1.11	0.00	0.56	1.00	6.65	1.03	1.04
C	1.12	1.73	1.63	3.06	2.00	1.70	0.83	1.35
hy	0.20	0.26	0.66	0.40	0.30	1.65	0.35	0.48
mt	.....	.....	0.70	.....	.....	.....	0.70	.....
hm	0.96	0.64	0.96	0.96	2.24	3.20	0.96	1.44
il	.....	.....	.....	0.34	0.36	0.46	.....	.....
ru	.....	.....	.....	.....	.....	0.14	.....	.....

* *See* Table 5 for name and location of samples.

In thin section the texture appears to be porphyritic with microgranular groundmass of somewhat irregular grain size, although it may be pyroclastic. Flow structure is locally prominent, especially around phenocrysts. Phenocrysts range up to 2 mm long.

Albite (An_3) makes up 60 per cent of the phenocrysts. It is unzoned, twinned, and moderately to considerably sericitized. About 25 per cent of the phenocrysts are quartz. They are highly corroded and have strained extinction.

Microperthite, low in albite, accounts for 10 per cent of the phenocrysts. It is considerably fresher than the albite although filled with minute inclusions.

A few hornblende phenocrysts are present but are largely replaced by iron ore, sericite, and piedmontite. The groundmass consists of quartz, feldspars, iron ore, epidote, and sericite.

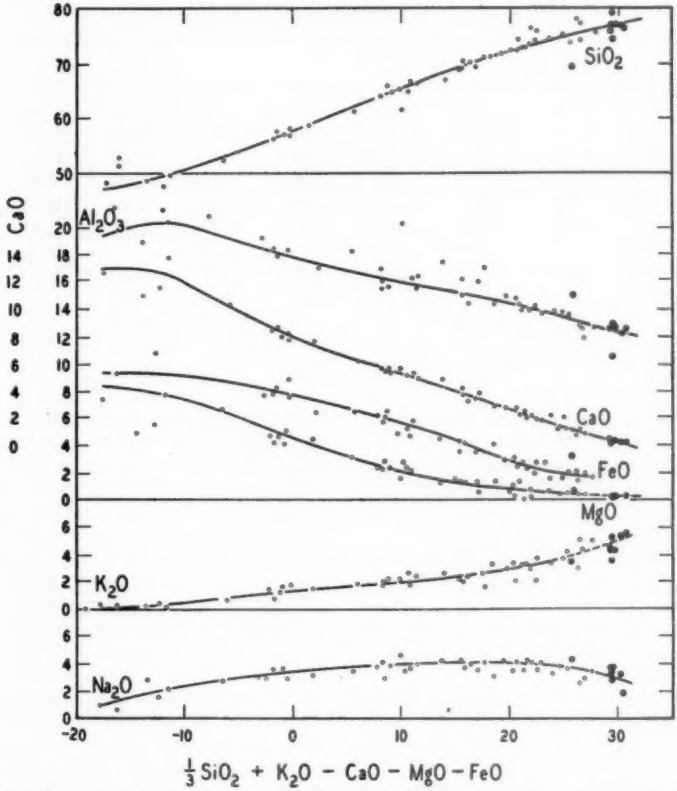

FIGURE 3.—VARIATION DIAGRAM FOR POWAY RHYOLITIC ROCKS AND ROCKS OF THE SOUTHERN CALIFORNIA BATHOLITH

In part after Larsen (1948). Open circles are batholith rocks; solid circles are the Poway Conglomerate rocks.

PC-1, SODA RHYOLITE TUFF: Megascopically this rock is massive, brownish purple, hard, and fresh. Phenocrysts of pink potassium feldspar and gray quartz 1–2 mm across compose nearly one-third of the rock. In a few places light-yellow plagioclase phenocrysts are visible.

In thin section the texture is seen to be pyroclastic with microgranular matrix composed of anhedral quartz and potassium feldspar. Irregular patches of the groundmass are more coarse-grained than most of the remainder.

Quartz occurs as resorbed crystals and irregular grains with sharp extinction as well as

some nearly euhedral crystals. It makes up approximately 50 per cent of the rock. Microperthite crystals amounting to 30 per cent are nearly euhedral, rectangular crystals charged with unidentifiable inclusions. Some alteration to sericite is apparent. Small, unzoned crystals of plagioclase (An_{5-10}) make up nearly 10 per cent. Crystals are euhedral to subhedral; some are fractured, and all are crowded with inclusions of sericite.

Iron ore is widespread but small in total percentage.

PC-19, SODA RHYOLITE TUFF: This is a light-gray, porphyritic rock in hand specimen. It has conspicuous white and gray phenocrysts which lack clear-cut outlines and blend into the aphanitic groundmass. Irregular fractures are numerous.

The microscopic texture is pyroclastic. Crystals 1 to 2 mm across make up 30 per cent of the rock. The finer matrix is microgranular and shows no flow features, although sheared zones are common; some are healed by recrystallization.

Approximately 50 per cent of the larger crystals are quartz; most show strained extinction or granulation or both. Many grains have clear centers surrounded by an outer, inclusion-free zone, which suggests two stages of growth. About half of the groundmass consists of quartz.

Albite (An_2) composes 30 per cent of the crystals. It is unzoned, twinned, moderately saussuritized, and occurs as crystals 0.1–1 mm long.

Microperthite (15 per cent) is slightly altered and contains less than 10 per cent albite.

Biotites, largely replaced by iron ore and muscovite, are sparsely distributed.

ORIGIN OF THE POWAY CONGLOMERATE

Former Theories

Though the origin of Poway Conglomerate is now largely an academic problem, at one time it was more important because of the gold which was being panned from the gravel near Ballena. Since gold was found in bed rock to the east, at Julian, the men who panned the gold concluded that a large river flowed westward or southward (toward the San Diego River) through the region and deposited the Poway Conglomerate (Fairbanks, 1893).

Later Ellis and Lee (1919) offered two hypotheses concerning the origin of the Poway Conglomerate. Either an arm of the sea extended eastward and coarse detritus accumulated in the trough, or the delta of a river covered the area.

Dusenbury (1932) suggested that a rejuvenated mountain mass to the east produced a great fan, the western end of which entered the sea to form a marine deltaic deposit.

Miller (1935) implied that the Ballena gravels and the Poway gravels are part of one large well-dissected river deposit. He stated that "... a wide ancient river valley once extended from north of Santa Ysabel southwestward past Witch Creek, Ballena, and the San Vicente Valley region."

Stock (1937), followed by Hertlein and Grant (1944), on the basis of the vertebrate fauna (which includes various mammals) and the character of the sediments (which resemble the non-marine Sespe), believed that the Poway Conglomerate is essentially continental in origin.

Objections to Former Theories

These explanations and hypotheses of origin must be supplemented, because they are preoccupied with only three points: the present outcropping pattern and/or the trace of gold and/or the fossils.

Because the outcrops are in a long, disconnected string from Witch Creek to the main body near Poway, this is not necessarily the only original area of deposition. The outcrops are now ridge tops capped by conglomerate, which is resistant. They could just as reasonably be the remnants of a large mass several hundred square miles in area that has been stripped away by erosion in connection with the uplift of the Peninsular Range.

No definite trend of size distribution can be found. There may exist a possible northeast-southwest range from abundant coarse gravel in the La Mesa area to less coarse, then coarse again north of Miramar, and again less coarse. This may be an accidental distribution of gravel. The size gradation found in river deposits or large fans is not apparent. Boulders between 1 and 2 feet in diameter were found almost everywhere in the conglomerate.

The trace of gold is found in the Ballena placer workings near Ramona in parts of Sections 17, 18, 19, 20, and 21 of T. 13 S., R. 2 E., San Bernardino Base and Meridian (Hertlein and Grant, 1944). The fact that gold is now found in appreciable quantities only around Julian does not mean that in Eocene time the gold-bearing rocks were limited to the Julian

area. Since no gold is known in the area to the west (most of which is now beneath the sea), the source of the conglomerate is more reasonably placed to the east.

The fossils perhaps indicate marine origin for some parts of the conglomerate and nonmarine for others. Certainly a major part of the main body is marine, on the basis of the microscopic and megascopic marine fossils. The vertebrate material could be deposited in a marine delta or bay. However, possibly this part of the formation is continental and interfingers with a marine facies.

An important point which has been overlooked is the composition of the conglomerate clasts, which are predominantly rhyolitic porphyry and tuff. Any explanation of the Poway Conglomerate must account for the source and abundance of this rock.

Other Rhyolitic Rocks of Southern California

Before attempting to offer any source for the clasts of the Poway Conglomerate it may be best to describe other rhyolitic rocks of Southern California that may have some relation to this problem. Rhyolitic rocks are relatively rare in Southern California, and in the immediate vicinity of the Poway outcrops such rocks are absent. In general, to qualify as a possible source for the Poway, a rock should be pre-Eocene and roughly within 50 miles of the present Poway outcrops (unless a special explanation for a more distant source can be presented).

However, rocks that qualify in the time element and roughly in character are present at distances of 100–125 miles from the main Poway Conglomerate outcrops. The writers do not presume that these are the source of the Poway rocks, but their physical mode of occurrence and petrographic associations are similar to those of the primary phase of the Poway volcanic series.

The closest group is 75 miles to the north, in the San Bernardino Mountains north of Big Bear Lake, where a series of volcanic rocks of possible Triassic age have been described by J. F. Richmond (1954, Ph.D. Thesis, Stanford Univ., Stanford, Calif.). The absolute age of these rocks has not yet been determined but assigned largely on the basis of intrusive relations of related rocks.

The San Bernardino Mountain rock nearest the rhyolitic type is a quartz latite porphyry which occurs in two facies. One shows a banding caused by flow lines, and the other is massive. These distinctions are of structural nature only. The composition of the rocks is identical. Richmond's description is as follows:

> "Allotriomorphic resorbed feldspar phenocrysts, as much as 3 mm. long, make up from 5 to 10 per cent of the rock. About half show albite twinning. Their composition ranges from oligoclase to albite. . . . Quartz phenocrysts are rare . . . Quartz in angular fragments as much as 0.2 mm. in diameter, may form half the groundmass. The remainder . . . is a cryptocrystalline aggregate of fine-grained quartz, sericite, and saussuritized feldspar."

Some quartz latite tuffs are associated with this rock in the field.

Two points in Richmond's work are of considerable interest in connection with the Poway study: (1) Though the two rocks are not identical they are roughly similar, and possibly they were much more alike before metamorphism of the San Bernardino Group (Richmond, personal communication). The second point is the mode of occurrence. In the San Bernardino Mountains three rock types occur in close association: a biotite-quartz monzonite, a granite porphyry, and the finer-grained porphyry complex which is predominantly quartz latite porphyry. These three rocks show gradational contacts which suggest that the porphyry complex is the outer rim of the batholith mass. The porphyry is dated as Triassic (?); if the indicated relation is correct, the peripheral material or initial phases of such a complex would probably be of a somewhat earlier age than the main batholithic mass, which is given as Cretaceous. This same relation may have occurred in the Peninsular Range between the granitic rocks and the rhyolitic material.

The Sidewinder volcanic series on the Mojave Desert in the area between Victorville and Barstow bears some petrographic relation to the Poway group. This was described in detail by Bowen (1954) in a report on the Barstow quadrangle. The Sidewinder Series is predominantly pyroclastic, and consists of interfingered and overlapped masses of dacite, quartz latite, latite-andesite, and rhyolite. The rhyolite and rhyolite tuffs are the rocks most closely related to the Poway group. Bowen refers to these as metavitrophyres and describes them as chertlike porphyries and dense felsites. The phenocrysts consist of perthitic orthoclase, oligoclase (An_{18}), and quartz. The groundmass is microcrystalline and is composed largely of potassium, feldspar, and quartz. The average

composition of phenocrysts and groundmass is given (Bowen, 1954, p. 46) as follows: quartz 25 per cent, olgioclase 18 per cent, perthitic orthoclase 12 per cent; ferromagnesian 5 per cent, and indeterminate (primarily orthoclase) 40 per cent. Bowen described gradations between crystalline rhyolitic varieties and definite pyroclastics. Felsites showing microcrystalline quartz and feldspar are common. The ferromagnesian minerals are usually altered to magnetite and hematite. Many of the specimens contains piedmontite in replacement clots and disseminated in the rock. Piedmontite is also a prominent characteristic of many of the Poway cobbles. The age of this series is again not definite. It is assigned to the Triassic (?) on the basis of the relation to the associated rocks. Bowen did not describe any relation of gradational contact or facies change. The Sidewinder Series is intruded by the Jurassic-Cretaceous granitic rocks.

Another, though minor, occurrence of rhyolitic rocks is found in the Santa Ana Mountains, 40–50 miles north.

The Santiago Peak Volcanics, named by Larsen (1948), are the youngest pre-Upper Cretaceous rocks in the area. Because these varied volcanic rocks are younger than the Triassic and intruded in the south by the quartz diorite of the batholith, they have been called Jurassic. There are many local variations in the Santiago Peak Volcanics. A very mild metamorphism has not obscured the bedded layers of different textures and composition. The volcanic rocks include tuffs, breccias, and flows. They are primarily andesitic, but there are some quartz latites (or dacite) and some rare rhyolites. The outcrops are nearly all severely weathered. Most of the rhyolites are nearly white or light green; the green is probably due to a secondary epidotization. They occur in several localities, notably in Tin Mine Canyon south of Corona, where a greenish weathered outcrop appears, and in Silverado Canyon where an almost white rhyolite intersects the road. Rock at the latter locality has less than 10 per cent phenocrysts in a near-white aphanitic groundmass. Most of phenocrysts are quartz and are at least five times more abundant than the pinkish feldspars. The outcrops of these rocks do not clearly show whether they are flows or intrusions. The narrow width of outcrop, 20 feet more or less, and the general relation to the rest of the formation suggests intrusive dikes. These rhyolites are unlike the Poway Conglomerate rhyolitic porphyries in general appearance.

A careful search of the exposed outcrop of the Santiago Peak Volcanics in the northern portion of the Santa Ana Mountains does not reveal any of the Poway rhyolitic material. Several geologists have mapped and given lithological reports of sections of the Santiago Peak Volcanics (Larsen, 1948; Schoellhamer, Kinney, Yerkes, and Vedder, 1954), but none has reported a rhyolitic rock of more than incidental significance, nor has anyone described a porphyry of the Poway type. A similar formation is found in San Diego County, largely in the San Luis Rey quadrangle. This was called the Black Mountain Volcanics by Hanna (1926a, p. 199). Later Larsen (1948), p. 23) assigned these rocks to the Santiago Peak Volcanics. There is some doubt that these are equivalent formations. However, the Black Mountain Volcanics in the San Diego area contain some material that may be related to the Poway porphyries and the welded tuffs. Work by Ellis E. Roberts and his San Diego State College students, in the San Ysidro Mountains near the Mexican border, between Lower Otay Reservoir and Dulzura, shows the presence of fine-grained siliceous igneous rocks, intruded by the plutonic rocks of the southern California batholith. Roberts (1957, unpublished) reports that "The most abundant rock unit is the pyroclastic material which grades from meta volcanic breccia to tuff and from dacite-andesite to rhyolite in composition. The second most abundant unit is a meta rhyodacite which may be intrusive or may be an extrusive." The red pebbles of the Poway are not evident, but the black and gray "porphyries" of the Black Mountain series are closely similar to Poway clasts. It may be that Black Mountain-Santiago Peak volcanic rocks are part of some volcanic activity preliminary to the emplacement of the main batholithic mass.

Proposed Hypothesis

In attempting to determine the source of the Poway Conglomerate, two points must first be established An areal limit must be determined for the source, and a time limit for their emplacement before their erosion.

The Poway Conglomerate is largely a shallow-water deposit and shows no signs of density-current deposition. It is not possible to set exact limits for the distance this material has been transported. To determine this the major element to be considered is the attrition from bedrock to present location. The data in Table 3 indicate that approximately 9 per cent of the

pebbles are 3 inches or more in their longest diameter; the largest are about 2 feet. Studies in the Alps by Fugger and Kastner (1895) indicate that, after about 10 miles of preliminary transport by various agencies, 1-foot blocks of granite and gneiss were moved 41 miles farther along the Salzach and reduced to 3 inches in diameter. On this basis an approximation for the possible distance of the source of the Poway pebbles may be given as 50 miles. This should be modified by two points which may offset each other. The rhyolitic material of the Poway that produces the majority of the large clasts is more resistant to fracture or abrasion than are granites. This would add some extra mileage to the total distance. However, the direction of travel of the pebbles was probably not a straight line. The reworking of the pebbles and some of the marine action may have added distance to the total mileage traveled but not in distance from the source. The writers therefore suggest that a circle 50 miles in diameter, drawn with its center at Poway just north of the largest mass of the conglomerate and near the center of its distribution, should encompass the source of the conglomerate.

This area reaches into the Pacific Ocean to the west, as far north as San Juan Capistrano and Elsinore, to the east to Borrego Springs, just east of the Peninsular Range, and south to a point about 20 miles beyond the Mexican border, a total area of nearly 8000 square miles. One-third of this, for which bedrock data are lacking, is under the Pacific Ocean; this leaves a little more than 5000 square miles, largely in San Diego County, as a source area that may be investigated for the rhyolitic Poway rock. This entire area has not been mapped in detail, but much of it has. Within these areal limits the typical Poway "porphyries" have not been found in the bedrock. Even by dredging along the coast north of La Jolla and on the shelf north of La Jolla no evidence of Poway rhyolitic rocks in place has been established. Only rounded pebbles washed out to sea from the land have been found (Emery and Shepard, 1945.) If any of the original red rhyolitic bedrock is present today within the probable areal limits of the Poway Conglomerate source, it must be a very small outcrop.

The displacement of source rocks beyond the 50-mile limit by faulting or folding is not considered a possibility, because faults known to have large lateral displacements are absent, as are appreciably folded rocks. The Elsinore fault, which crosses the northeastern part of the source area, shows no major displacements in the bedrock (Larsen, 1948, Pl. 1). The possible major lateral displacements on the San Andreas fault and San Jacinto fault have no bearing on the Poway problem, since the outcrop and probable source area are both to the west of the faults.

With regard to the time of primary occurrence of the sodic rhyolitic Poway suite, the Poway Conglomerate may not be their first appearance as detritus material. The Upper Cretaceous Trabuco and Baker formations contain some quartz porphyry pebbles. These formations are exposed along the west and southwest slope of the Santa Ana Mountains and southward sporadically along the Peninsular Range to the San Diego area. The Trabuco Formation is a well-weathered, poorly sorted, nonfossiliferous, coarse conglomerate. It may be nonmarine, or a basal conglomerate which becomes marine in the upper zones. The overlying marine zone is called the Baker Member of the Ladd Formation. The conglomeratic zones of these two formations are very similar. Popenoe (1941), in a pebble-count study, reports that in the Trabuco he finds that in some samples more than 80 per cent of the pebbles are of an andesitic type and as much as 5 per cent of "rhyolitic" material. In the Baker-Trabuco conglomerates the quantity of these "rhyolitic" pebbles ranges from less than 1 per cent to a high of 30 per cent. All the rhyolitic material cannot be identified as the Poway type, but some of the pebbles are identical in the hand specimens. The dark-gray "porphyries" are common. However, the red pebbles so typical of the Poway are very rare. Thin sections of material collected and studied by us indicate that some of the dark-gray quartz-bearing porphyries are almost identical to some of the Poway pebbles; even the welded tuff texture is present. This occurrence at least suggests that it is possible that the primary sources of the Poway rhyolitic pebbles were already present and were starting to be eroded during the Cretaceous.

It becomes obvious after a study of the primary rhyolitic outcrops of the southwestern corner of Southern California that, with the exception of some parts of the Black Mountain Volcanics, no direct source for the sodic-rhyolitic Poway suite can be found. The Black Mountain Volcanics probably contributed some of the Poway pebbles. The source of the typical red pebbles of the conglomerate is still unknown. It is probable that the red phase of the volcanic rocks was a localized occurrence, perhaps a secondary nature. Therefore, regardless

of the time of occurrence, the primary source has been destroyed or obscured in some manner.

In view of the necessary conditions of time and place, the writers suggest that the most likely position for the rhyolitic material was on or in the roof rock of the Peninsular Range batholith, which qualifies in the respect of having been completely removed by erosion within the proper areal limits and time limits. The rhyolitic suite may have been an independent igneous activity, or it may have been directly related to the batholith emplacement. If the former is true, a localized volcanic activity took place some time near the final phases of batholith activity or at the time of its unroofing. However, it would be an unusual coincidence for two igneous activities in the same area and very close in time to be unrelated. Moreover, the chemical characteristics of the rhyolitic Poway pebbles conform closely to what would be expected if the assumed relation to the batholithic magma is correct. Therefore, the writers propose that there is a probable relation between the sodic-rhyolitic Poway suite and the composite granitic batholith of the Peninsular Range.

Resumé of the Geologic History of the Poway Conglomerate

The work of E. S. Larsen, Jr., is the starting point in attempting a hypothetical geological history of the events associated with the igneous intrusions of the west San Diego County region and the deposition of the Poway conglomerates. The sequence of intrusion of the Southern California batholith is given by Larsen (1948, p. 137) as follows: "The gabbros were intruded first, then the tonalites, then the granodiorites and finally granites". He also adds that "For the most part, the older intrusions were nearly or entirely crystalline when the next intrusive was emplaced."

This sets the stage for the events accompanying the intrusion. From some of the bedrock evidence available in the region, Larsen (1948) assumes a normal intrusion of the batholith by stoping and other processes into the Paleozoic metasediments and early Mesozoic rocks (Bedford Canyon and some of the Santiago Peak-Black Mountain formations) found in the area. There is little or no evidence to support any granitization of any of these rocks.

Toward the end of the magmatic differentiation process in the batholith magma, an acid magma (about 75 per cent SiO_2) had been produced. The most salic portion was probably concentrated at the roof which, by this time, was rather thin and weak. Some of this magma escaped through the roof in minor local intrusions as dikes and sills which led to the surface to form several small flows with large masses of interbedded pyroclastics. The latter were of a very fine-textured type similar to a fine-grained welded tuff. The extrusions were slightly different in physical appearance and mineral composition but were essentially the same chemically. Some flows were almost free of phenocrysts and formed many thin sheets crumpling the lower layers as the new ones formed. Though the groundmasses were of various colors, nearly all were porcelaneous or chertlike. Local secondary effects may have produced the red facies which include piedmontite and/or iron ore. The areas affected by these rhyolitic extrusions were small and not evenly distributed. One extrusive center was probably in the general region of Ramona. Probably several centers of eruption occurred. All this extrusive action probably took place contemporaneously and early enough to permit some phases of the batholith to invade it later. The final emplacement of the granite phase of the composite batholith was completed in Lower-Upper Cretaceous time.

The uplift of the area initiated the erosion that stripped the cover from the batholith. Most of the roof rocks were metamorphic rocks, covered in part by the andesitic volcanic rocks of the Santiago Peak and Black Mountain formations and the rhyolitic extrusive rocks. The detritus produced from the metamorphic rocks, except for the quartzites and pure quartz, was not, in general, very durable. The andesitic detritus was most abundant because of the extensive original exposure of the andesitic rocks. Though the rhyolitic rocks were very resistant, their smaller original volume could show only a small per cent in the presence of the large quantity of andesitic rocks. This relation is shown in the Cretaceous conglomerates, such as the Trabuco and Baker formations, where the small quantity of plutonic rocks indicates little erosion of the batholith, but the high per cent of andesitic rock and the small per cent of rhyolitic material indicates the relative abundance of these rocks that were being eroded from the roof of the batholith. These two formations may not be typical of all the deposits of that period; others may have had much more of the rhyolitic material, or even less.

Subsequent reworking of this material, possibly several times, to account for its wide-

spread and uniform character, had a tendency to concentrate the hard, chertlike, rhyolitic rocks. As long as new eroded material was not added, the rhyolitic material gained in percentage as the other rocks were reduced to sands or chemically decomposed.

The Poway Conglomerate was then spread in a large sheet, in places 1000 feet thick, over an area of 400–500 square miles. Some of the formation was marine, and probably the margins were continental. Subsequent events had little effect on the Poway Formation, which was uplifted and dissected in the Pleistocene to produce the present situation.

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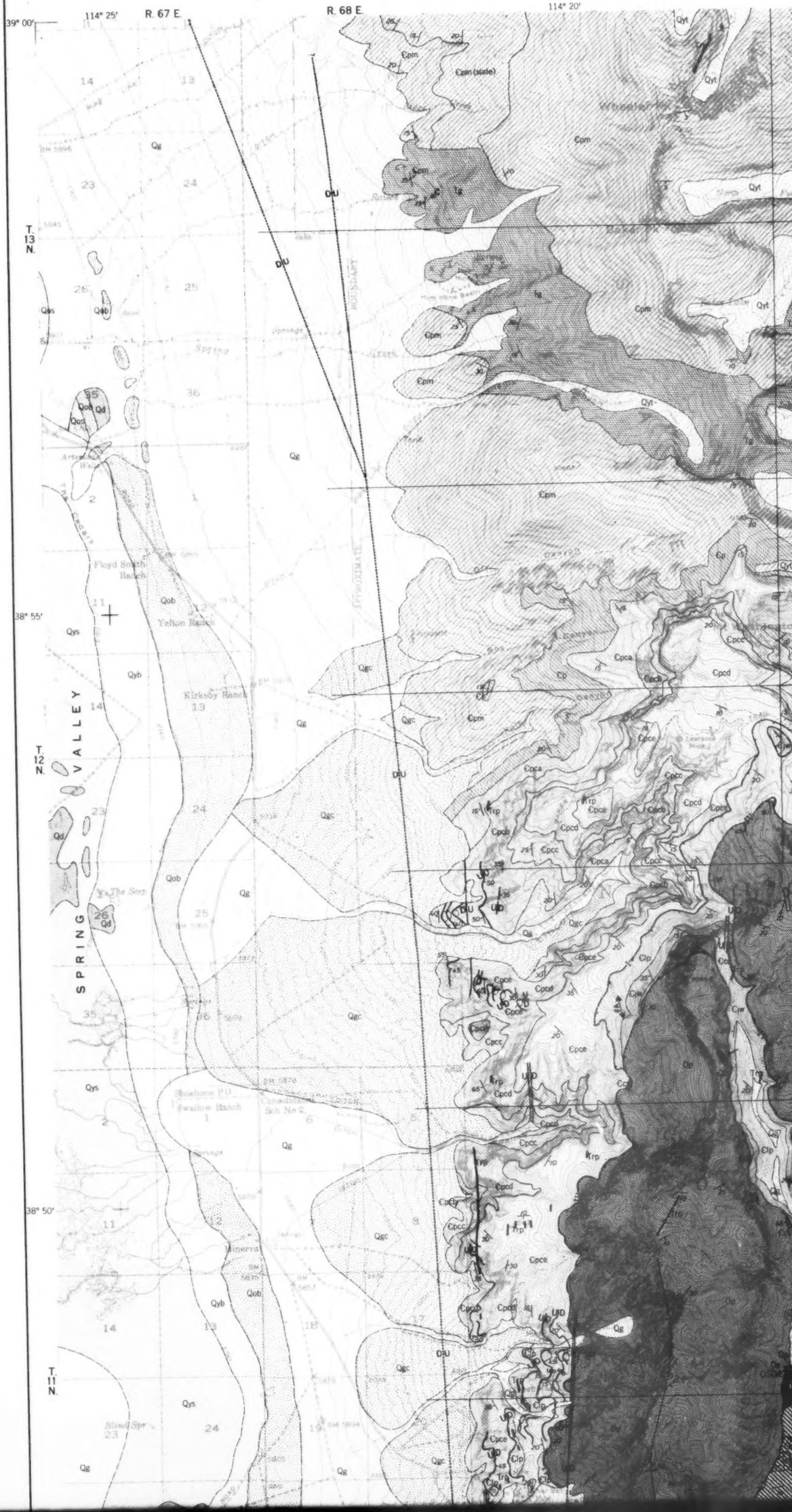
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Swallow Ranch
Wheeler Pk.
Qyt
Qpm
Cpm (state)
Qg
Qgc
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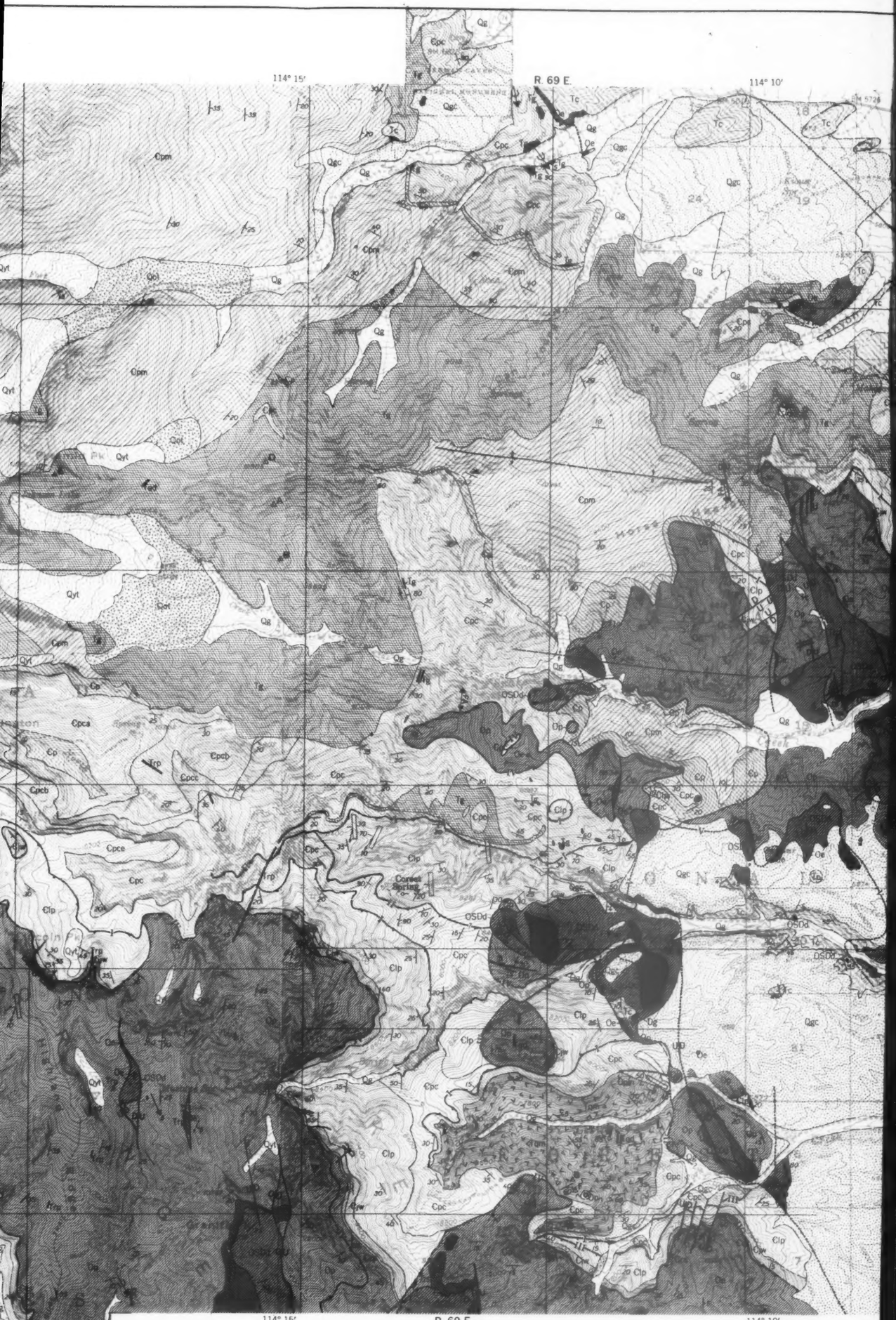

EXPLANATION

SEDIMENTARY ROCKS

ALLUVIAL

Qg — Gravel

Qgc — High level gravel and conglomerate

GLACIAL

Qyt — Younger till

Qot — Older till

LACUSTRINE

Qyb — Younger beach gravel

Qys — Younger silt and clay

Qob — Older beach gravel

Qos — Older silt and clay

AEOLIAN

Qd — Dune sand and silt

ANGULAR UNCONFORMITY

Tc — Conglomerate

ANGULAR UNCONFORMITY

Shale, limestone and sandstone

Guilmette formation(?)

Fish Haven, Laketown, Sevy and Simonson dolomites undifferentiated

QUATERNARY

TERTIARY(?)

DEVONIAN

ORDOVICIAN

MISS. AND PENN.(?)

DEVONIAN

IGNE

Tqd — Quartz diorite dike

Trp — Rhyolite porphyry dike

MA

CONTACTS

Exposed

Approximately located, inferred or gradational

FAULTS

SEDIMENTARY ROCKS

ALLUVIAL

Qg — Gravel

Qgc — High level gravel and conglomerate

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ANGULAR UNCONFORMITY

Tc — Conglomerate

ANGULAR UNCONFORMITY

Shale, limestone and sandstone

Guilmette formation(?)

Middle Ordovician

Fish Haven, Laketown, Sevy and Simonson dolomites undifferentiated

POSSIBLE DISCONFORMITY

EXPLANATION

AEOLIAN

Qd — Dune sand and silt

QUATERNARY

IGNEOUS ROCKS

TERTIARY (?) TERTIARY OR QUATERNARY

- Tqd — Quartz diorite dike
- QTd — Dacite porphyry flow
- Trp — Rhyolite porphyry dike
- Tg — Granodiorite stock
- Tqm — Quartz monzonite porphyry stock

MAP SYMBOLS

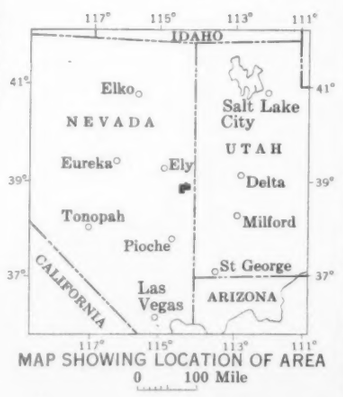
CONTACTS

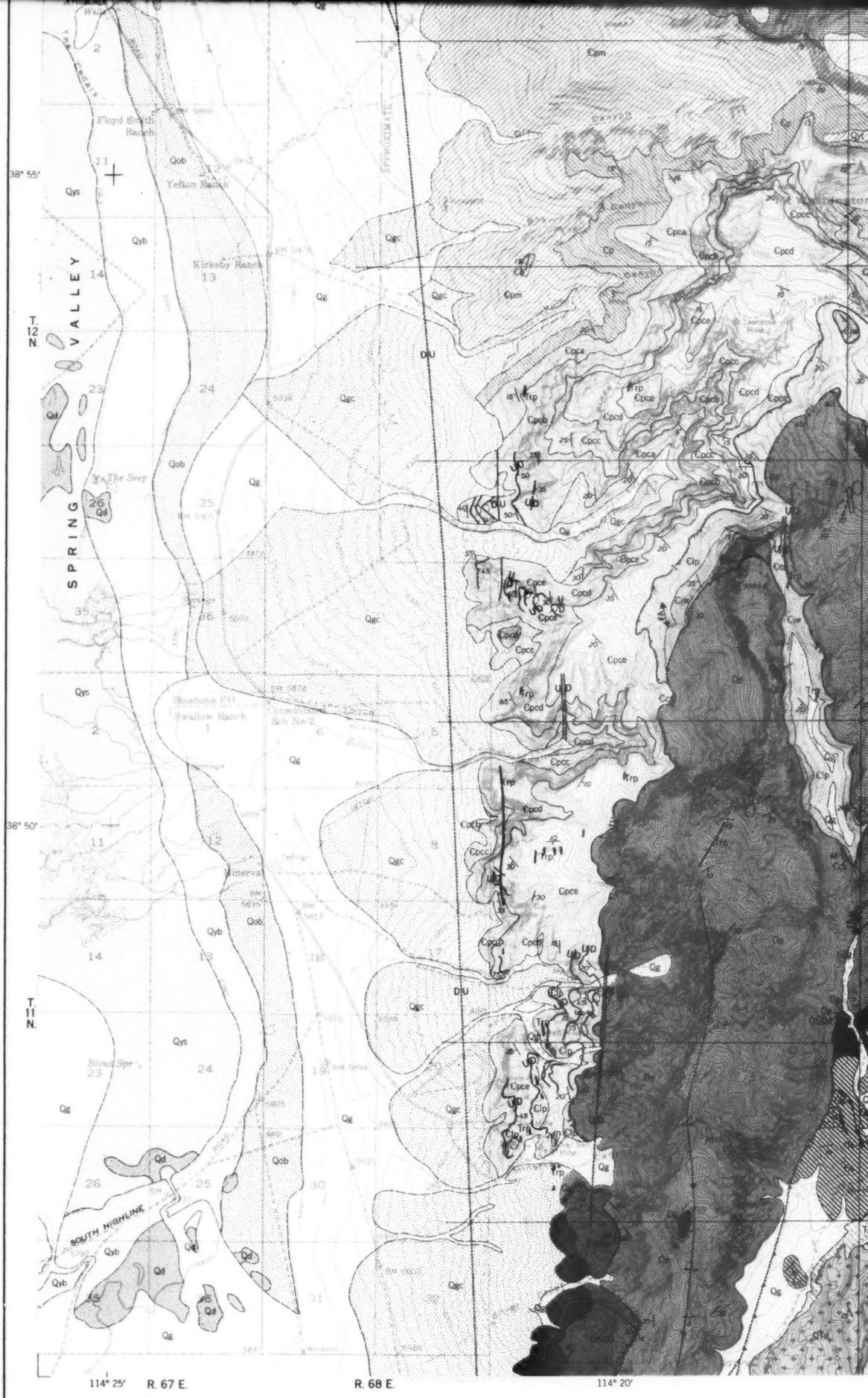
- Exposed
- Approximately located, inferred or gradational

FAULTS

FOLDS

- Trace of anticlinal axial plane showing plunge direction
- Trace of synclinal axial plane showing plunge direction
- Horizontal fold axis

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Kirksby Ranch
Swallow Ranch
Shoshone PO
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Minerva
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38° 55'
38° 50'
T. 12 N.
T. 11 N.
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R. 68 E.
114° 20'
Qob
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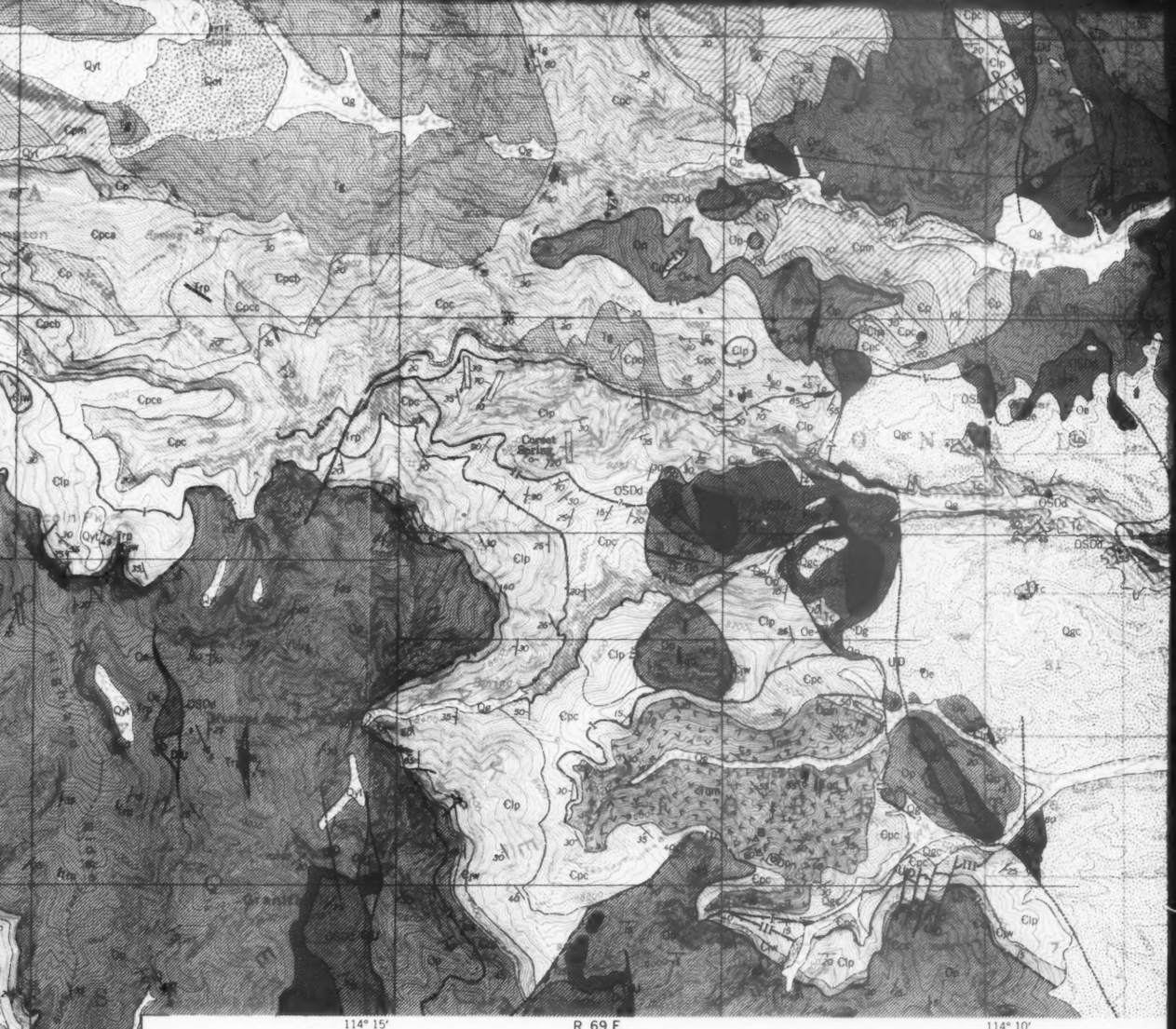

EXPLANATION

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Qys — Younger silt and clay

Qos — Older silt and clay

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QUATERNARY

ANGULAR UNCONFORMITY

Tc — Conglomerate

TERTIARY (?)

ANGULAR UNCONFORMITY

Shale, limestone and sandstone

Guilmette formation(?)

Fish Haven, Laketown, Sevy and Simonson dolomites undifferentiated

DEVONIAN

POSSIBLE DISCONFORMITY

ORDOVICIAN TO DEVONIAN

MISS. AND PENN.(?)

Eureka quartzite — *Middle Ordovician*

Op — Pogonip group — *Lower Ordovician*

ORDOVICIAN

Ccs — Corset Spring shale

Cjw — Johns Wash limestone

Clp — Lincoln Peak formation — *Upper Cambrian*

Cpc — Pole Canyon limestone — *Middle Cambrian*

Cpce, Cpcd, Cpcc, Cpcb, Cpca

Members a through e distinguished locally.

Cp — Pioche shale — *Lower Cambrian*

Cpm — Prospect Mountain quartzite

CAMBRIAN

IGNEOUS ROCKS

Tqd — Quartz diorite dike

Trp — Rhyolite porphyry dike

MAP

CONTACTS

Exposed

Approximately located, inferred or gradational

FAULTS

Exposed, high-angle fault

U, *upthrown:* D, *downthrown.*

Approximately located or inferred

Probable fault

Thrust fault

Teeth on upper plate side.

Concealed fault

30 — Dip of fault plane and strike slip movement

A, B — Collection sites referred to in text

OF THE SOUTHERN SNAKE RANGE, NEVADA

By

Harald Drewes

Scale 1:63,360

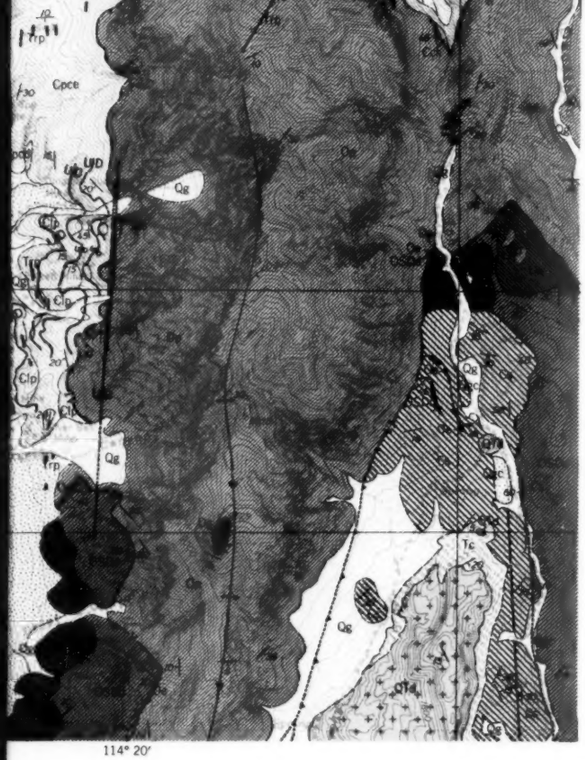
1 ½ 0 1 Mile

1 .5 0 1 2 Kilometers

Contour interval 40 feet

Datum is mean sea level

1957



SEDIMENTARY ROCKS

ALLUVIAL

Qg
Gravel

Qgc
High level gravel and conglomerate

GLACIAL

Qyt
Younger till

Qot
Older till

LACUSTRINE

Qyb
Younger beach gravel

Qob
Older beach gravel

Qys
Younger silt

Qos
Older silt and

ANGULAR UNCONFORMITY

Tc
Conglomerate

ANGULAR UNCONFORMITY

Shale, limestone and sandstone

Guilmette formation(?)

Fish Haven, Laketown, Sevy and Simonson dolomites und

POSSIBLE DISCONFORMITY

Eureka quartzite

Pogonip group

Middle Ordovician
Lower Ordovician
Upper Cambrian

Ccs
Corset Spring shale

Cjw
Johns Wash limestone

Clp
Lincoln Peak formation

Middle Cambrian

Cpc: Cpce, Cpcd, Cpcc, Cpcb, Cpca
Pole Canyon limestone
Members a through e distinguished locally.

Lower Cambrian

Cp
Pioche shale

Cpm
Prospect Mountain quartzite

GEOLOGY OF THE SOUTHERN SNAKE R

By

Harald Drewes

Scale 1:63,360

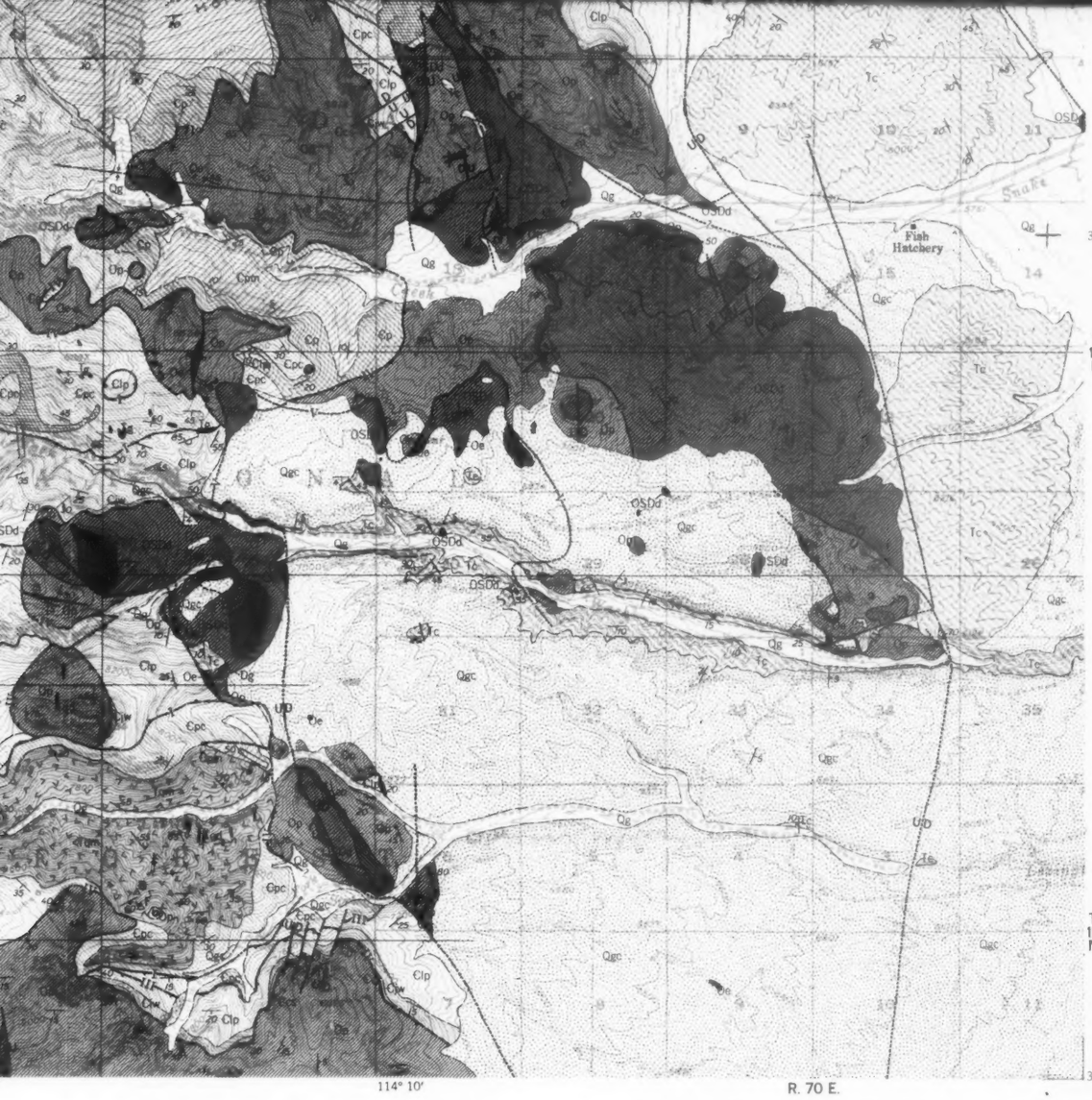
1 ½ 0 1 Mile

1 .5 0 1 2 Kilometers

Contour interval 40 feet

Datum is mean sea level

1957



EXPLANATION

AEOLIAN

Qd — Dune sand and silt

Qys — r silt and clay

Qos — lt and clay

s undifferentiated

ly.

QUATERNARY

TERTIARY (?)

ORDOVICIAN TO DEVONIAN

MISS. AND PENN. (?)

CAMBRIAN

IGNEOUS ROCKS

TERTIARY (?) TERTIARY OR QUATERNARY

Tqd — Quartz diorite dike

Trp — Rhyolite porphyry dike

Qtd — Dacite porphyry flow

Tg — Granodiorite stock

Tqm — Quartz monzonite porphyry stock

MAP SYMBOLS

CONTACTS

Exposed

Approximately located, inferred or gradational

FAULTS

U D — Exposed, high-angle fault
U, *upthrown:* D, *downthrown.*

Approximately located or inferred

Probable fault

Thrust fault
Teeth on upper plate side.

Concealed fault

30 — Dip of fault plane and strike slip movement

△B — Collection sites referred to in text

FOLDS

Trace of anticlinal axial plane showing plunge direction

Trace of synclinal axial plane showing plunge direction

FA — Horizontal fold axis of minor anticline

Plunge of minor anticline

Attitude of fold with inclined axial plane showing dip and bearing and plunge of axis, where applicable

BEDDING AND FOLIATION

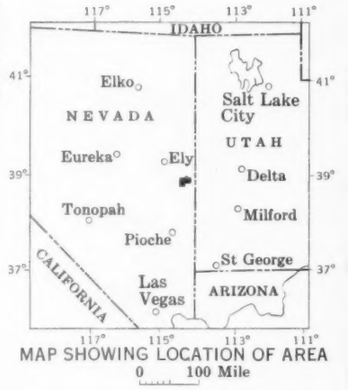
30 — Strike and dip of beds

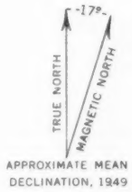
90 — Strike of vertical beds

70 — Strike and dip of overturned beds

70 — Strike and dip of foliation

Strike of vertical foliation





Topographic base used:
U.S.G.S. quadrangles Wheeler Peak, Nevada, and Garrison, Nevada-Utah; 1949-1950.

Geology mapped in 1952-1953.

RANGE, NEVADA

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 69, PP. 221-240, 4 FIGS., 2 PLS. FEBRUARY 1958

STRUCTURAL GEOLOGY OF THE SOUTHERN SNAKE RANGE, NEVADA

BY HARALD DREWES

ABSTRACT

At least 14,000 feet of Cambrian to Pennsylvanian(?) strata are exposed in the Wheeler Peak and Garrison quadrangles which straddle the southern Snake Range in east-central Nevada.

Six thrust faults, not all of which have had much displacement, generally follow the shaly formations, and in places some of the thrust faults cut out several thousand feet of strata. The upper thrust plates moved relatively east-northeastward. A fence diagram and a structure contour map of one of these thrust faults show the coincidence of some of the troughs or structurally low belts in the thrust plane with areas of gouged-out underlying shaly beds. The sequence of truncation of the thrust plates indicates that successively higher plates were active as the underlying ones were essentially locked in place. Some normal faults in the widespread thrust breccia sheets are restricted to one or several thrust plates and are believed to be contemporaneous with the thrust faults. Only one major orogeny of probably late Mesozoic or early Tertiary age is necessary to explain these structures; very likely they are the near-surface expression of one major thrust zone.

Two Tertiary granitic stocks that show little contact alteration are located in areas of regional and local structural domes and have drag features along their borders that indicate a viscous magma was forcibly injected at relatively low temperature.

Several periods of movement along high-angle faults bordering the range are recorded by major unconformities in the probably late Tertiary to Recent deposits flanking the range and by various erosion surfaces.

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Introduction

Location

The Snake Range of Nevada lies near the western boundary of Utah and is about equidistant from the southern and northern boundaries of that state (Fig. 1). A northern extension of the range swings into Utah and is known as the Deep Creek Range; to the south the Wilson Creek Range forms a low continuation. This entire geographic unit stretches over 150 miles between Pioche, Nevada, and Gold Hill, Utah. Midway between these towns and about 50 miles southeast of Ely, Nevada, is the southern block of the Snake Range proper. The U. S. Geological Survey Wheeler Peak and Garrison quadrangles, which were used as base maps, straddle the central portion of the southern Snake Range.

Scope of Work

During the summers of 1952 and 1953 the writer spent more than 100 days mapping about 250 square miles of the Wheeler Peak and Garrison quadrangles. Part of the time he had an assistant, but even so this project can be considered no more than a detailed reconnaissance study. This structural mapping project constitutes a partial fulfillment of the requirements for the degree of Doctor of Philosophy granted by Yale University in 1954.

Acknowledgments

The author is grateful for the financial assistance of the U. S. Government through the G. I. Bill, and to the Stanolind Oil Company for a grant. Further assistance was rendered by C. R. Longwell, supervisor of this dissertation, A. R. Palmer who helped in measuring some of the Cambrian rocks, and Harry M. Gehman and Bernard Drewes, who assisted in the field. To these and many others whose constructive criticism has served to germinate ideas and to guide their development, the author extends sincere thanks.

Physiography

The southern Snake Range is a bold northward-trending range that has an asymmetrical transverse profile; the drainage divide is nearer the west flank. South of Lincoln Peak the major drainage is divided between the eastward- and westward-flowing streams and those flowing southward. Many peaks along the divide are 11,000 feet high or higher; Wheeler Peak (also known as Jeff Davis Peak in some early accounts) is 13,063 feet high.

The two flanks of the range differ in many respects. Foothills are absent along the west flank; the front of the range on that side drops off abruptly to the gentle slopes of alluvial fans, which are only slightly incised by water courses. The fans on the west flank grade into the floor of Spring Valley, where beaches and spits, indicating the past existence of a lake, and sand dunes constitute the chief relief features. The east flank, on the other hand, is fringed by a belt of gently rolling hills that is widest between Big Wash and Snake Creek. Gentle alluvial slopes descend eastward beyond the edge of the map to Lake Creek and northeastward to an arm of the Bonneville Basin. Big Wash and Lexington Creek follow deep valleys in the younger alluvial deposits for several miles east of the foot of the mountains.

Paleozoic Stratigraphy

General Summary

The southern Snake Range stratigraphic sequence consists of at least 14,000 feet of rocks of Early Cambrian to Pennsylvanian age (Fig. 2). Only minor and probably local stratigraphic breaks are recognized in the deposi-

tional history of this section, most of which consists of carbonate rocks. Most post-Paleozoic sedimentation is confined to coarse clastics of unknown thickness and uncertain age, though possibly late Tertiary to Recent.

and massive limestone is also common to the regional picture, but the correlation of the carbonates of Middle and Upper Cambrian age remains difficult. Because of this difficulty many stratigraphic names used in this region

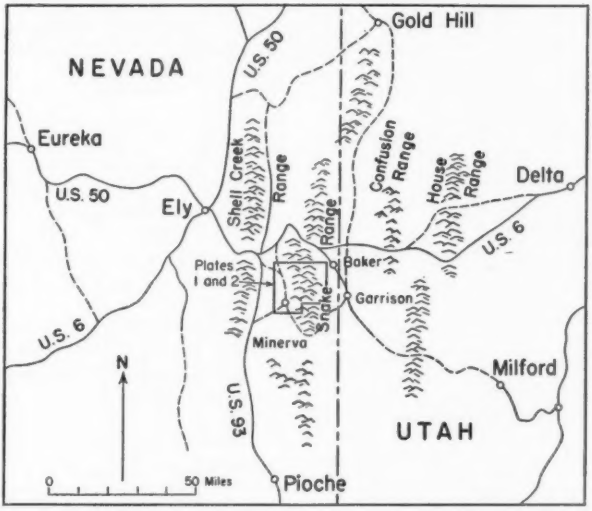

FIGURE 1.—INDEX MAP SHOWING AREA WHICH COVERS PARTS OF THE WHEELER PEAK AND GARRISON QUADRANGLES IN NEVADA

Many stratigraphic details remain obscure because of the structural complexity of the area. The information is most complete for 4000 feet of Middle and Upper Cambrian strata. The measured sections in this part of the column are presented by Drewes and Palmer (1957). Full sections of the thick Prospect Mountain quartzite and Pogonip group have not been located, and partial sections cannot be correlated adequately because of the numerous bedding-plane faults and the scarcity of key beds. Information regarding the late Paleozoic formations is augmented by data from near-by sections, as they occur only in a few scattered, much faulted areas. Even the sections of Middle and Upper Cambrian strata are separated by a bedding-plane thrust fault. Therefore, the thicknesses of some formations presented here are only tentative.

The major features of this stratigraphic sequence are very similar to others of eastern Nevada and western Utah. Cambrian(?) sedimentation begins with the remarkably uniform, thick Prospect Mountain quartzite. The overlying succession of thin Pioche shale are only local. The Ordovician sequence of thick, cherty limestone of the Pogonip group, Eureka quartzite, and dark-brown dolomite is also widespread, and lithologic correlation is relatively reliable. Massive gray dolomite continues upward into the Middle Devonian and is overlain by Middle and Upper Devonian limestone. Fine clastics reappear in the Upper Devonian rocks and continue upward in the Carboniferous as shale and some sandstone interbedded with limestone.

Prospect Mountain Quartzite

The Prospect Mountain quartzite exposed here is similar to that at Eureka, Nevada, defined by Hague (1883, p. 254; 1892, p. 34). It differs little from the basal Cambrian clastic rocks in the eastern and central part of the Great Basin. Most of this quartzite is restricted to the northern half of the map area where it covers 25 square miles. Slopes underlain by this quartzite commonly are veneered with quartzite blocks and are broken by scattered small cliffs. At higher altitudes tongues of

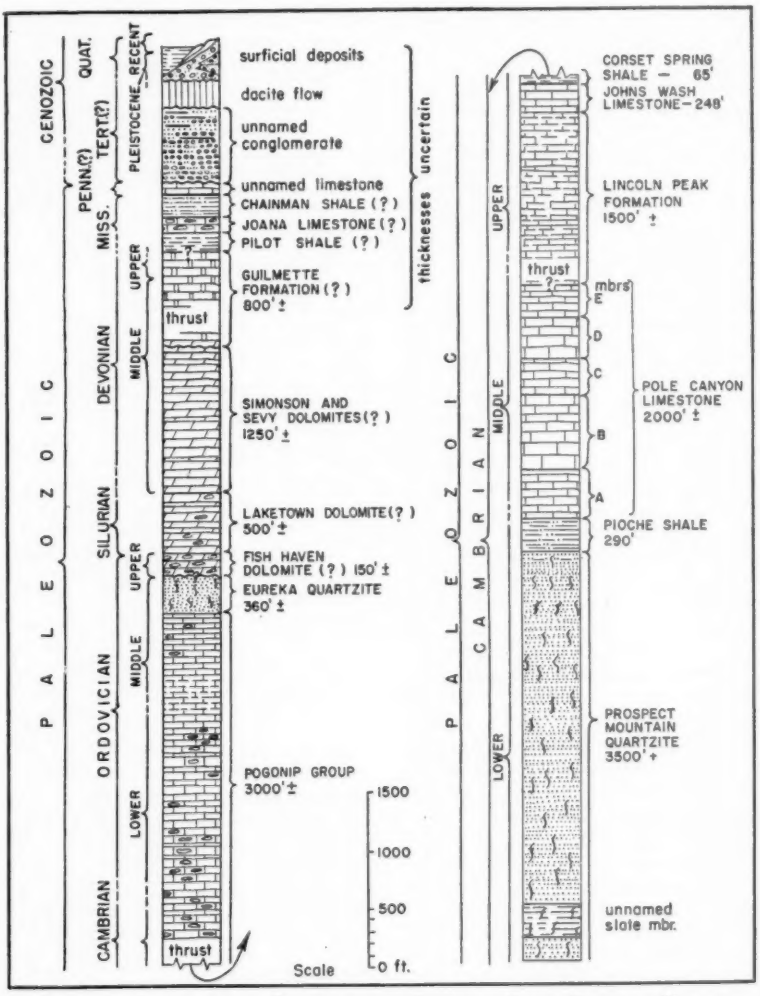

FIGURE 2.—STRATIGRAPHIC COLUMN OF THE SOUTHERN SNAKE RANGE IN NEVADA

The Pilot shale (?), Joana limestone (?), Chainman shale (?), and overlying unnamed limestone are mapped as one unit; similarly the Fish Haven dolomite (?), Laketown dolomite (?), Simonson dolomite (?), and Sevy dolomite (?) are mapped as one unit.

rock-creep debris are conspicuous. The base of this formation is not exposed, and the top appears to grade into the Pioche shale by means of more abundant and thicker shaly beds.

The purity, uniformity, and massiveness of the Prospect Mountain quartzite are its most striking features. Most of the grains are medium-sized quartz sand, but a few grains of muscovite, amphibole, and plagioclase are recognized. Locally the quartzite contains lenses of granules and small quartz pebbles and flat, shaly chips. Bedding averages about 3 feet in thickness, but some beds are as much as 6 feet thick. Cross-bedding is common, and the foreset beds generally face westward. Scour marks and oscillation ripples are present in places. The thick platy beds are cut by regu-

larly spaced joints, which help to provide the coarse blocky rubble that veneers slopes underlain by this formation. Fresh surfaces of the quartzite are characteristically light gray,[1] but pale brown and tints of green, red, purple, and yellow are also common.

In thin section the grains appear subrounded, are partly interlocking to sutured, and are presumably parallel to the bedding. Cementing material consists of small quartz crystals, totals less than 5 per cent of the volume of most rocks, and is absent in others. Where this cementing quartz is missing the interlocking, sutured texture is best developed, and this suggests that at least some of the quartz cement is locally derived. Only rarely are shadowy outlines of detrital grains discernible within their sheathing of optically continuous quartz. Minor quantities of sericite appear as veinlets and patches.

These intercalated shaly and phyllitic units which resemble the overlying Pioche shale are progressively more common upward in the quartzite. In the Hub Mine Basin on the west side of the range and in the upper Lehman Creek (north of the map) areas there is a slate member several hundred feet thick. This slate unit may be similar to one of those in the Deep Creek Range described by Nolan (1935, p. 6) as part of the Prospect Mountain quartzite. This slate member is dark purplish gray and greenish gray and weathers brownish to greenish black. Most, but not all cleavage in the slate parallels bedding.

The exposed thickness of this quartzite is estimated at 3500 feet. Although fossils (of Middle Cambrian age) first occur several hundred feet above the Prospect Mountain quartzite, the lithologic similarity between this quartzite and the Prospect Mountain quartzite at its type locality, and its conformable relation to the overlying Pioche shale north of Mount Washington provide the present basis of correlation. However, other recent work in the northern Snake Range by J. C. Hazzard and Peter Misch (Personal communication) indicates that there are one or more older, more metamorphosed quartzite units in the area. But until more of the range is mapped all the quartzite in the Wheeler Peak and Garrison quadrangles is considered Prospect Mountain quartzite.

[1] Color descriptions, wherever practical, follow the Rock Color Chart of the National Research Council, 1948. Number designations are used only for names that occur more than once in the chart.

Pioche Shale

The Pioche shale, first defined by Walcott (1908, p. 11) and later redefined by Westgate and Knopf (1932, p. 8–10), conformably overlies the Prospect Mountain quartzite. This formation makes a thin outcrop band across the range from Pole Canyon to the ridge south of Snake Creek. Slopes underlain by this shale form a bench between the cliffs of adjacent limestone and quartzite.

The argillaceous beds in the Prospect Mountain quartzite are more abundant upward and are transitional to the basal Pioche shale. Much of this shale is hard, quartzitic, and micaceous. Several thin limestone units are interbedded near the base. Bedding commonly is 2–3 inches thick, and many bedding planes contain anastomosing organic markings. The shale is dusky red to dusky brown and olive black and measures 290 feet in thickness near Mount Washington.

Useful fossils have not been found in this formation in this area, but the stratigraphic relations to adjacent formations are sufficiently similar to known sections to suggest the correlation with Pioche shale of Early or Early and Middle Cambrian age.

Pole Canyon Limestone

The Pole Canyon limestone, about 2000 feet thick, is a massive cliff-making formation of thick alternating gray and white units most of which are devoid of textural and structural variations. The limestone is defined and described in greater detail by Drewes and Palmer (1957, p. 110–113). In the Wheeler Peak quadrangle 5 members, *A* through *E* in ascending order, have been mapped, but farther east structural complexity obscures the simple color variation, and the limestone has not been subdivided.

The basal limestone member, *A*, 415 feet thick, and the medial limestone member, *C*, 320 feet thick, are finely crystalline, light- to dark-gray units containing calcite blebs and *Girvanella*-like forms. The medial member contains fine laminae and oolites and also fragments of *Glossopleura* sp. and *Chancia*.[2]

Members *B* and *D*, bracketing the medial gray limestone, form white cliffs on the flanks of Mount Washington. Only a few gray beds indicate attitudes; other bedding and sedimentary features are absent. These two

[2] Cambrian fossils were identified by A. R. Palmer, U. S. Geological Survey.

members have a moderately coarse saccharoidal texture and consist of calcite crystals that disaggregate readily. Member *B* is 630 feet thick; member *D* is 220–380 feet thick. No fossils were obtained from them.

The topmost member, *E*, is 200–300 feet thick and is more variable and less recrystallized. Some flat-pebble intraformational conglomerate and finely laminated cross-bedding were observed. Locally bright-red and yellow-brown silty partings occur, and toward the top of the member such shales become dominant. Fossils common in this limestone are *Peronopsis* sp., *Parkaspis* sp., *Ptychagnostus Triplagnostus* sp., and *Tonkinella* sp., all of Middle Cambrian age. The Pole Canyon limestone resembles the Howell limestone, Dome limestone, and Swasey formation in the House Range, Utah (Deiss, 1938), except that the shaly zones characteristic of the Howell limestone and Swasey formation was absent.

The textural alteration of most of the Pole Canyon limestone is metamorphic and preceded the thrust-fault movement, for unaltered Pogonip is faulted against altered Pioche shale along lower Snake Creek and against lower Pole Canyon limestone south of the North Fork of Big Wash. If this interpretation is accepted, and if the age of the plutons, described below, is accepted as younger than the thrust faults, then the metamorphism of part of the Cambrian rocks is not a contact effect of these plutons. Either an older contact metamorphism or perhaps even regional metamorphism must be considered as alternate explanations.

Lincoln Peak Formation

The shale and shaly limestone of the Lincoln Peak formation of Middle and Late Cambrian age, defined by Drewes and Palmer (1957, p. 113–114), everywhere appears faulted over the Pole Canyon limestone. A thickness greater than 1000 feet has been measured, but structural interpretation east of the section at Lincoln Peak may require twice that thickness.

Shaly limestone beds, generally less than 2 inches thick, are separated by siliceous and micaceous shale beds. Fresh surfaces are medium gray, whereas weathered ones are light olive gray to pale yellow brown. Grayish-red colors are most common 800 feet from the top of the formation, and fossiliferous limy nodules are common toward the top of it. Structural complexities are responsible for the stratigraphic thinning and thickening of this formation and hence for the interruptions of its characteristic bench topography.

The lowest faunal horizon, the *Elrathina* zone, is equivalent in age to Albertan. Above are successively the *Cedaria-Tricrepecephalus* fauna correlated with that at the base of the Dresbachian, and the *Aphelaspis* fauna (probably equivalent to the Franconian). The base of the rocks equivalent to the Trempealeau is about 250 feet above the top of the Lincoln Peak formation. This formation is, in part, lithologically similar to the Weeks limestone of the House Range, Utah (Wheeler and Steele, 1951), the Hicks formation of the Gold Hill district, Utah (Nolan, 1935) and the Secret Canyon shale in the Eureka district, Nevada (Wheeler and Lemmon, 1939). However, these units occupy only portions of the stratigraphic interval represented by the Lincoln Peak formation and are not correlative among themselves.

Johns Wash Limestone

The Johns Wash limestone, defined by Drewes and Palmer (1957, p. 115), conformably overlies the Lincoln Peak formation and is 248 feet thick. Chert-free crystalline limestone of various textures characterizes this formation. The basal half of the formation is relatively uniform, finely crystalline, and clastic and has coarse cross-bedding and several conspicuous beds of edgewise limestone-pebble conglomerate near the bottom. Above this unit is a wavy-bedded, platy limestone containing calcite blebs and a few oolitic beds. The topmost unit is clastic and coarsely crystalline.

The fauna collected near the top of the Johns Wash limestone is meager and not sufficiently diagnostic for precise dating; it includes *Pterocephalia* cf. *P. occidens* Walcott, *Ditremacephalus* sp., and *Pseudagnostus* sp. Part or all of this limestone is tentatively considered equivalent to basal Trempealeau; possibly some rocks equivalent to the Franconian are present toward the base of the formation.

Corset Spring Shale

The olive-gray shale that forms a small, persistent bench above the Johns Wash limestone is defined by Drewes and Palmer (1957, p. 115–116) as the Corset Spring shale. It is probably stratigraphically conformable with the underlying limestone and has a maximum thickness of 65 feet. Above a medial limestone this pencil-fractured, clay-shale contains fossilif-

erous limy nodules. An abundant trilobite assemblage that includes *Linnarasonella girtyi* Walcott, *Dunderbergia* sp., *Berkeia* sp., *Elvinia* sp. and *Housia* sp. indicates an early Trempealeau age for the Corset Spring shale. This formation is probably correlative of part of the Dunderberg shale of the Eureka district, Nevada, and of part of the Mendha formation of the Pioche district, Nevada (Wheeler and Lemmon, 1939).

Pogonip Group

About 3000 feet of limestone of the Pogonip group here correlated with the unit as originally described by Hague (1892) conformably overlies the Corset Spring shale. Subdivision of the Pogonip group has not yet proved to be practical in this range, for much of the limestone is uniform and low-angle faults commonly disrupt its continuity. The base of this group is conformable with, and grades into, the Corset Spring shale. As used in this paper, the Pogonip group includes at least 550 feet of Upper Cambrian rocks which are impractical to separate from the lithologically similar Ordovician portion.

The Pogonip group is characteristically a calcilutite but contains bioclastic and coarse crystalline beds as well. Thin partings of pale-red to pale-yellow-brown silty shale commonly separate platy limestone beds whose average thickness is between 2 and 4 feet. Chert nodules and beds are widespread in much of the formation but constitute less than 5 per cent of any unit. High in the Pogonip group are olive-gray silty shales containing interbedded limestone. Locally coquinoid limestone and limestone conglomerate are also found.

In much of the Pogonip group fossils are not abundant, but collections are adequate to show the presence of fossils similar to those of the high Trempealeau, Canadian, and Chazy assemblages. Faunal zones G2, J, and M, as described by Ross (1949; 1951) and by Hintze (1952) are best represented.

Eureka Quartzite

About 360 feet of light-colored quartzite of Middle Ordovician age overlies the Pogonip group and underlies the dolomite of Late Ordovician age. This quartzite is generally brecciated and is distributed in scattered knobs rather than as a continuous belt. Still, it forms here, as elsewhere in the Basin and Range area, a useful lithologic marker.

This formation is generally a pure quartzite; in places it is a quartz sandstone with minor amounts of carbonate cement. Where relatively undisturbed, the Eureka quartzite is massive or has a poorly defined bedded structure and has a uniform texture of moderately well-rounded, medium-sized sand grains. Where brecciated the siliceous matrix contains sufficient ferruginous matter to color the quartzite pale reddish brown to moderate red.

Bedding-plane thrust faults may explain much or all of the patchy distribution of the Eureka quartzite; whereas evidence for a possible local unconformity, if it ever was present, may be hopelessly obliterated in this area. This problem parallels that of the discontinuous outcrop band of Upper Cambrian shales except that regional data failed to suggest any unconformities in the Upper Cambrian rocks, whereas there is a distinct possibility of one associated with the Eureka quartzite as shown by Ferguson (1933), Sharp (1942), and Kay (1951). No fossils have been found in this quartzite; it is generally accepted as a Middle Ordovician formation.

Undifferentiated Dolomites of Ordovician, Silurian and Devonian Ages

For practical reasons the 2500–3000 feet of dolomite of Late Ordovician, Silurian, and Early Devonian age is mapped as one large unit. Only a few square miles of these rocks are present, generally as klippen, and probably represent various parts of the Fish Haven, Laketown, Sevy, and Simonson dolomites found in the Gold Hill area (Nolan, 1935).

The Fish Haven dolomite in the southern Snake Range has a thickness of 150–200 feet and consists of brownish-gray crystalline dolomite with dark chert nodules, corals, crinoid stem fragments, and casts of brachiopods and gastropods. The overlying Laketown dolomite contains alternating medium-gray and brownish-gray coarsely crystalline dolomite. Locally the thicker cliff-making darker beds contain corals and chert and produce a fetid odor when struck. More rarely a coral-reef lithology and fauna are found. The dimorphous variety of *Halysites* and *Virgiana* are indicative of a Silurian age. The next overlying 250–300 feet is a medium-gray finely crystalline dolomite with beds 4–18 inches thick, and the uppermost 1000–1200 feet of dolomite is finer textured, lighter gray and more regular in bedding. Together these dolomite units are probably equivalent to the Sevy dolomite and

Simonson dolomite. The thicknesses for these formations are obtained from R. Rush (Personal communication) who measured a section near Big Spring, south of the map area.

Guilmette Formation(?)

Several small thrust slices and klippen contain a greenish-gray limestone. No one area contains more than a few hundred feet of this limestone, and not all areas contain the same fauna; the total thickness is unknown. Most of the Guilmette formation(?) is aphanitic to finely crystalline and has a fetid odor when struck. Bedding planes are somewhat irregular and beds are about 3 inches thick. *Spirifer argentarius*, *Martinia*, and possibly *Atrypa* point to a Middle Devonian age, whereas *Cyrtospirifer whitneyi* indicates Late Devonian age.

A tentative correlation of this limestone with the Guilmette formation at Gold Hill (Nolan, 1935) is suggested.

Mississippian and Pennsylvanian(?) Rocks

In the Johns Wash—Murphy Wash area shale, limestone, and minor amounts of sandstone containing a Carboniferous fauna overlie the Guilmette formation(?). The Carboniferous units consist of the following ascending sequence: olive-black clay-shale and siltstone containing nodules and coquinoid limestone; several hundred feet of brownish-gray to purplish-gray, cherty, crinoidal limestone; possibly more red to dark gray, fissile shale; and light purplish-gray limestone and several ferruginous sandstone beds containing productid casts and a large *Calamites*. J. S. Williams and H. M. Duncan of the U. S. Geological Survey identified the meager collection of Spirifers, Dictyoclostids, corals, bryozoans and gastropods as Carboniferous, probably upper Mississippian and possibly some lower Mississippian and lower Pennsylvanian. The lithologic sequence and the fauna suggest a possible correlation with the Pilot shale, Joana limestone, and Chainman shale of the Ely, Nevada, area.

Thrust Faults

General Statement

Thrust faults produce many of the stratigraphic discontinuities in the southern Snake Range. Shale and thin-bedded shaly limestone between more massive units localize most of these faults. Some thrust faults have a relatively large displacement, others have much less, and one of them need only be a bedding-plane slip. The six thrust faults and bedding-plane slips form a belt which crosses the range with a northeasterly trend. In this belt omission of strata, probably by piecemeal grinding away, is typical; in only one place are beds repeated by a thrust fault. The plates above some of these thrust faults moved relatively east-northeastward and, as only one epidsode of thrust faulting is postulated, the movement direction of all plates is inferred to be about east-northeastward. Some faults truncate underlying ones so that movement on the lower ones must have ceased when movement on the upper ones began. Minor jostling of some plates produce normal faults that are restricted to those plates.

The inferred sequential development of the bedding plane slips and thrust faults is shown diagrammatically in Figure 4. All combinations of thrust plates shown in section *G* are mapped on Plate 1. Since the thrust faults and bedding plane slips are so nearly parallel to the bedding they are numbered consecutively from the base upward. This numerical sequence corresponds, in part at least, to a time sequence.

Similar thrust faults are reported from adjacent ranges, but bedding plane slips are generally not mentioned. The thrust faults described at Gold Hill, Utah, by Nolan (1935) are most like those in the southern Snake Range. More prominent thrust faults occur in the northern Snake Range (Darton, 1908; Hazzard, Misch, Wiese and Bishop, 1953), in the Shell Creek Range (Misch and Easton, 1954) and in the Pioche district (Westgate and Knopf, 1932). Some of these thrust faults are believed to be related to, or even continuous with the faults in the southern Snake Range.

Description of Faults

Thrust fault I follows the contact between the Pole Canyon limestone and the Lincoln Peak formation. Along most of its trace there are no signs of much displacement, but south of Lincoln Canyon and 2 miles east of Lincoln Peak the Lincoln Peak formation thins over a short distance from about 1500 feet to only a few hundred feet. This thinning must be caused by thrust I; the final wedging out of the Lincoln Peak formation at these places is accomplished by thrust fault IV. The loss of about 1000 feet of strata adjacent to a bedding plane fault and the absence of displaced blocks near by requires

considerable displacement—at a very minimum a displacement in the order of 1000 feet. There is no indication, however, that fault I is a thrust fault of very large displacement. About a mile northwest of the confluence of the North and South Forks of Big Wash the trace of fault I rises on the northern valley wall, where it is marked by sheared, phyllitic debris and moderately contorted, but rarely overturned, overlying beds. Drag folds along the North Fork of Big Wash have limbs exceeding 10 feet and axial planes that strike northward and dip steeply westward. To the west the trace of this thrust fault gradually rises on the southern valley wall and is cut by thrust fault III, which repeats thrust fault I and the adjacent rocks.

Above fault III in the plate the trace of fault I runs from 1 mile southeast of Lincoln Canyon to 1 mile east of Lincoln Peak, where thrust fault IV truncates it for a mile. Near Lincoln Canyon phyllitic shales are contorted into small drag folds with north-northwestward-trending axes. The trace of fault I follows around the north and east side of peak 11,064 and then along the South Fork of Big Wash it lies within the top of the Pole Canyon limestone. Similarly, around Lexington Creek the trace of fault I alternately follows the top of the Pole Canyon limestone or lies near the top of that formation. North of Snake Creek two short segments of a thrust fault at this horizon are inferred to be part of fault I.

Thrust fault II branches from fault I or truncates it along the south wall of the South Fork of Lexington Creek. The amount of thinning of the shales overlying fault II also suggests that the displacement is larger than that of a bedding slip, but perhaps not much larger. In this area it is possible that faults I and II are minor branches of just one thrust fault, fault I; the plate between them is not exposed over a large enough distance to show whether both bounding thrust faults have large displacement.

Thrust fault III, which repeats part of the Lincoln Peak formation, the Pole Canyon limestone, and a section of thrust fault I between these formations is exposed for only about 3 miles along the south wall of the North Fork of Big Wash. At its eastern end about 2 miles west of the junction of the North and South Forks of Big Wash the trace of fault III is covered by younger plates. At its western end the trace of fault III is mapped only a short distance into the Pole Canyon limestone about 1 mile north-northwest of peak 11,064. In this area are many large contorted fractures, west of which the location of the fault is uncertain.

Thrust fault IV is the most continuous one. It generally follows the base of the Pogonip group and brings it onto lower parts of the Corset Spring shale or any formation down to the Pioche shale. These tectonic transgressions of the plate overlying fault IV are gradual and of relatively small magnitude along Johns Wash and the South Fork of Lexington Creek; elsewhere, as east of Lincoln Peak, the transgression is abrupt, and many hundreds of feet of beds that normally are present below the Pogonip group are missing. In places east of Granite Peak and peak 11,064 the Pogonip group rests conformably on the Corset Spring shale, and thrust fault IV appears within the Pogonip group several hundred feet above its usual stratigraphic position. Here it is marked by an abrupt break in the attitude of the limestone. The rock immediately above fault IV is generally shattered and brecciated on the east flank of the range. Less commonly the underlying beds are upended, truncated, and dragged northeastward. The axes of most drag folds in the underlying shaly rocks trend northward to northwestward and in one place the axial plane is inclined westward.

Thrust fault V approximately follows the Eureka quartzite, which is generally brecciated and is faulted out in many places. In a few localities, as at Granite Peak, where this quartzite and the overlying dolomite are unbroken, the position of this fault is uncertain; perhaps it lies higher within the dolomite. About 2 miles south of Granite Peak and beyond the edge of the map the Eureka quartzite, about 360 feet thick, is faulted out along thrust fault V. The thin slice of Eureka quartzite, about 15 feet thick, between the two normal faults about a quarter of a mile northwest of Granite Peak is best explained by thrust faulting between two periods of normal faulting, for neither top nor bottom contact of the thin slice of quartzite is continuous with the same contact across the normal faults. But aside from the complications of normal faulting the important point is that more than 300 feet of quartzite are lost over a very short distance. Under these circumstances the thrust fault V is probably continuous but difficult to find in places, rather than a disharmonious bedding plane slip which has alternating zones of no displacement and zones of little displacement. Just how much displacement is involved is difficult to judge; an order of movement of

hundreds of feet is a bare minimum, thousands of feet far more likely, and even a few miles not unreasonable.

Thrust fault VI places the limestones of Middle and Upper Devonian and Mississippian age upon various parts of the dolomite of Lower Devonian, Silurian, and Upper Ordovician age and, on the ridge between Minerva Canyon and Johns Wash, upon the Pogonip group. As with fault V, because of its high stratigraphic position only klippen and small segments of the upper plate appear in the map area. Thrust fault VI gradually cuts out fault V along Johns Wash and is exposed in a stripped shear surface along the west side of that wash.

The geometric configuration of thrust plane IV is shown in Figure 3. This plane is gently arched from 7000 feet at Minerva to 11,000 feet along the divide of the range and down to 6500 feet along lower Baker Creek. The similarity of this arch to the stratigraphic doming about Wheeler Peak suggests a common origin postdating thrust movement. Superposed on this major arch are smaller eastward-trending anticlines and synclines, each 500–1000 feet in amplitude and several miles wide. These folds are sufficiently parallel to the tops of the stocks to suggest the same relationship. A northward-trending trough on this thrust plane reaches the surface near the upper fork of the North Fork of Big Wash, where the Lincoln Peak formation is missing. Possibly the missing 1800 feet or more of Corset Spring shale, Johns Wash limestone, Lincoln Peak Formation and the top of the Pole Canyon limestone was gouged out all along this trough.

Summary and Conclusions

A summary of the field observations shows that, although not every thrust fault has the same characteristics, they have an overall similarity and, hence, probably belong to one episode and environment of thrust faulting. The features believed to have some bearing on the development of these faults are shared by at least two faults, but not necessarily the same two faults for each characteristic. With the possible exception of fault IV, none requires much movement; perhaps the movement is in the order of thousands of feet rather than in miles. Drag folds occur in more than a dozen localities and are associated with faults I, III, IV, and V. Their axes trend from northeastward to north-northwestward, but most trend northward or northwestward. Four of the areas with drag folds have inclined axial planes, all of which dip westward. The direction of movement and movement sense on these four thrust faults (and presumably on the others, too) is such that the upper plates moved relatively east-northeastward.

Gouged-out zones in lower plates are common only beneath plates containing competent rocks and are associated with faults IV, V and possibly VI. Similarly, shattered features are associated with plates containing competent rocks, as those above faults IV and V. Where normal faults not associated with the much younger block faults occur in areas of thrust faults, they are restricted to one thrust plate. There is no evidence to indicate that episodes of thrust movement are separated by episodes of relative tension, as has been postulated by Nolan (1935) at Gold Hill.

In places faults IV, V, and possibly VI truncate underlying faults abruptly enough to suggest an age relation between these faults: The overlying ones are successively younger. Similarly, fault I is older than fault III where fault I is repeated, with the adjacent formations, northeast of peak 11,064. Fault III is inferred to be older than IV solely because this relationship follows the pattern established by the four other faults. Fault II may be older than, or of the same age as fault I; tentatively it, too, will be inferred to fit the pattern set by the other faults. The observed and inferred age relations are diagrammatically illustrated in Figure 4. These age relations are relative rather than absolute, and only one episode of compressive stress is needed to explain them. Apparently movement on the slightly older, lower thrust planes ceased when friction was sufficiently great on those planes to build up stresses on other, high potential planes.

The age of the thrust faulting remains vague. Rocks as young as Upper Devonian and possibly even as young as Pennsylvanian(?) are involved with these thrust faults. The granitic plutons are believed to postdate the faults but are themselves poorly dated, for Mesozoic and most or all Tertiary rocks are absent. If the age of the plutons is tentatively accepted as middle Tertiary, an age commonly accepted for such plutons in the eastern part of the Basin and Range, then the thrust faults may be Mesozoic or early Tertiary.

Intrusive Rocks

Granodiorite Stock

A large granodiorite body is exposed in and north of Snake Creek on the east flank of the

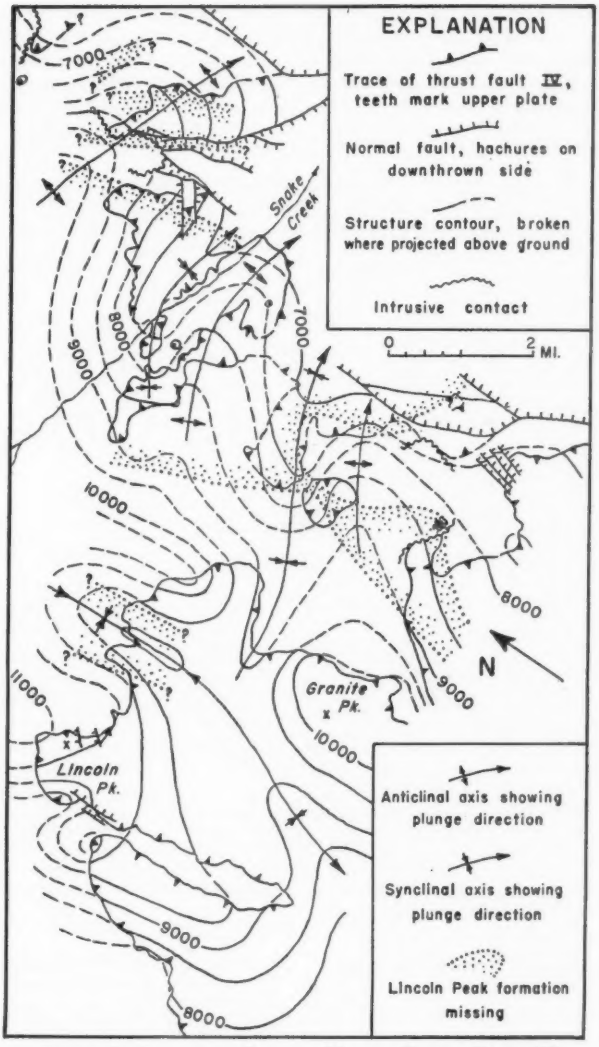

FIGURE 3.—STRUCTURE CONTOUR MAP OF THRUST PLANE IV

range and in Williams Canyon on the west flank. This body is essentially domelike; the eroded roof of the stock was probably not much higher than the present peaks in the granodiorite area. Toward the top of the stock, where the contact with the wall rocks is flat, it is essentially concordant. Along the outer margin of the dome, where this contact is steep or moderate, the relations with the enclosing Prospect Mountain quartzite, Pioche shale, and Pole Canyon limestone are discordant. Locally along steep contacts the host rock is dragged upward. The enclosing sedimentary rocks form a broad dome about this pluton.

The internal structure of the stock is fairly uniform. Weak foliation is barely discernible along some borders. Joints are widely spaced, and one persistent set in the western part of the stock is vertical, strikes N. 60°–80° E., and is commonly followed by quartz veins. A few aplite and micropegmatite dikes cut the stock. Inclusions are rare or absent. The average

grain size of this light-gray granodiorite is one-eighth–one-fourth of an inch. A zone of fine-grained granitic rock 5–10 feet wide borders part of the southern end of the stock. More commonly the granodiorite is aplitic near the margin of the stock, and micropegmatitic apophyses cut the near-by host rock. The texture of the rock generally is hypidiomorphic granular.

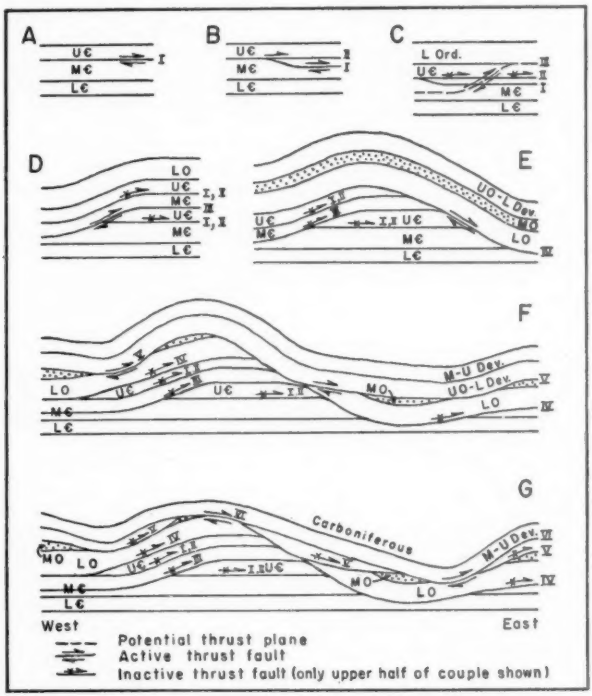

FIGURE 4.—DIAGRAMMATIC SECTIONS OF THE POST-PENNSYLVANIAN (?) SEQUENCE OF THRUST FAULTS AND BEDDING PLANE SLIPS

Sections *A* through *G* show their progressive development. From section *D* through *G*, faults I and II, which lie close together, are shown as one.

In Tables 1 and 2 are summarized the mineralogical and chemical composition of the granodiorite. Quartz, plagioclase, and microcline are the essential constituents; plagioclase is about twice as abundant as microcline. Biotite is the most persistent accessory mineral and sphene, pyroxene, muscovite, apatite, and magnetite-ilmenite are recognized.

The plagioclase ranges from sodic andesine nearest the center of the stock to oligoclase toward its margins and to albite in aplitic bodies. The anorthite content in zoned plagioclase ranges at least 8 per cent. Kaolinitic alteration beclouds the plagioclase, and sericite is abundant near quartz veins.

Microcline is slightly more abundant near the margin of the stock and is very abundant in the aplitic and micropegmatitic dikes. Most microcline is coarsely perthitic; patches and rims of clear albite are apparently exsolved from the potassium feldspar. The absence of alteration in the albite indicates that the alteration is strongly selective or that it precedes the suggested exsolution.

Biotite is commonly strongly pleochroic in deep reddish-brown and dark-brown colors indicative of a relatively high iron and titanium content. Refractive index values of Ny = 1.62 to 1.64 also indicate that this is a lepidomelane biotite with an estimated ratio Mg:Fe = 4:6. Toward the margin of the stock muscovite appears instead of, or with, the biotite. In computing the chemical composition of the rock

the average biotite composition of sodium granites shown by CIPW and modified by the determined Mg:Fe ratio is used.

The pyroxene is scarce and can be tentatively determined in only one specimen, A. The grains are equidimensional and have a perature. Larsen's value of 573° C, as presented by Hunt, may be a closer average for the temperature of intrusion of the granodiorite stock in the southern Snake Range.

Forceful intrusion most effectively provided room for the body by doming of the overlying

TABLE 1.—MINERALOGICAL COMPOSITION (MODE) OF GRANODIORITE FROM THE SOUTHERN SNAKE RANGE, NEVADA, IN WEIGHT PER CENT

Specimens A and B are from the Snake Creek Basin; C from the Hub Mine Basin; D from peak 10,503 south of Timber Creek; E from the spur east of Pyramid Peak. Precise locations are shown in Plate 1.

Mineral	A	B	C	D	E
Quartz	19.9	21.2	31.8	39.2	34.1
Plagioclase	50.7(An 27)	51.7(An 28)	41.0(An 28)	33.3(An 22)	39.7(An 17)
Microcline	20.4	19.1	16.8	17.1	24.4
Biotite	5.8	6.1	9.5	8.0	1.4
Sphene	0.3	0.5	0.3	t	—
Magnetite	t	0.2	0.6	—	0.4
Pyroxene	2.9	0.3	—	2.2	—
Apatite	—	t	—	0.2	—
Muscovite	—	0.9	—	—	—

birefringence of about 0.035–0.038, (−) 2V 75°–85°, ZAC 83°–87°, and Ny considerably over 1.705. The composition of this pyroxene is estimated to be acmite 60, diopside 20, hedenbergite 20 according to Winchell and Winchell (1951, p. 415).

Evidence of contact metasomatism is meager. Near the sill-like granitic tongue in Big Wash the limestone host contains abundant muscovite that may be related to the igneous body. A few traces of epidote and silica enrichment are also found. Slight textural changes of the country rock that may have been caused by the granodiorite emplacement are probably obscured by the effects of the earlier, low-grade metamorphism postulated above.

Magmatic intrusion at relatively low temperature seems likely. The meager evidence for a contact chilled zone and the relatively unchanged country rock suggests that at the present level of exposure the invading magma was relatively cool. In a somewhat similar case Hunt (1946, p. 17; 1953, p. 165) has shown that stocks and laccoliths of partly crystallized granitic magma can invade even poorly consolidated shale with only slight contact effects. Spencer is cited by Bateman (1950, p. 42) as showing that the exsolution of microcline requires a temperature of 400° C; if the observed exsolution is accepted as a deuteric reaction, then this could be a minimum intrusion temrocks and possibly also by dislocating some blocks in the roof area. Granitization, if at all effective, played a minor role. The relatively uniform composition and texture, the absence of regional alkali metasomatism, the presence of fine-grained border zones and small apophyses in the host, when considered together, favor a magmatic origin. The general similarity of the configuration of the top of the granodiorite to the top of the overlying Prospect Mountain quartzite (contoured but not presented here) suggests local upbowing in the response to forceful magmatic intrusion. The strongly discordant margins and local updrag of adjacent host rocks near Snake Creek (Pl. 2) and the apparent dislocation of the strata across this section suggest that, in part, block faulting of the roof of the stock somewhat akin to the "trapdoor" faulting of Knechtel (1944) accompanied the doming. Stoping, accompanied by assimilation and perhaps by foundering, is a mechanism necessary to explain the missing part of Prospect Mountain quartzite, Pioche shale, and less commonly also part of the Pole Canyon limestone.

The more silicic margins, the lack of a pronounced chill zone, and the very minor flow structure may, perhaps, be equally well explained by a granitization hypothesis, as has been argued by Stringham (1953) at Bingham Canyon, Utah. There a granitic stock is also

emplaced in a weakly metamorphosed quartzite. The extensive feldspathization of the quartzite that is common at Bingham Canyon, however, is absent in the southern Snake Range. Can two granitic bodies of seemingly the same petrographic province have such different origins? Or are these origins so different?

The age of this stock remains a matter of conjecture. The granodiorite is younger than the thrust faults which it truncates. The age of this body can not be effectively bracketed by stratigraphic units, as formations of definite Mesozoic or Tertiary age have not been recognized in this range. A middle Tertiary age is common to other stocks in the eastern part of the Basin and Range region; possibly this body has a similar age.

TABLE 2.—CHEMICAL COMPOSITION OF GRANODIORITE FROM THE SOUTHERN SNAKE RANGE, NEVADA, CALCULATED FROM THE MODE

Oxide	A	B	C	D	E
SiO_2	68.1	68.4	71.6	75.4	75.9
Al_2O_3	16.9	17.4	14.3	12.1	13.6
Fe_2O_3	0.9	0.5	0.8	0.4	0.3
FeO	1.0	1.0	1.6	1.5	0.4
MgO	0.9	0.9	1.2	1.4	0.2
CaO	3.2	3.1	2.3	1.8	1.4
Na_2O	4.6	4.4	3.6	3.1	3.9
K_2O	3.9	3.8	3.6	3.6	4.2
H_2O	0.2	0.2	0.4	0.3	0.1
TiO_2	0.3	0.3	0.4	0.3	t
P_2O_5	—	t	t	0.1	—
	100.0	100.0+	99.8+	100.0+	100.0+

Quartz Monzonite Stock

Moderately foliated, porphyritic quartz monzonite intrudes the Pole Canyon limestone along the north fork of Lexington Creek. The top of this stock is domal, and the contact of this body with the host rock dips steeply eastward and less steeply westward. Around the margin of this stock the foliation parallels the plane of contact, whereas in the center of the body the foliation is vertical and has a variable strike. At least some of the margins are fine-grained. Much of this rock is moderately sericitized.

Modal analyses of specimens relatively free of the sericitic alteration are listed in Tables 3 and 4. Estimates from other specimens show that rock G with the higher potassium content is the most representative. Phenocrysts consist largely of feldspars; the embayed borders of some of these suggest that possibly they are porphyroblasts. Plagioclase in the range of calcic oligoclase to sodic andesine is about as abundant as microcline. Accessory minerals include biotite, muscovite, apatite, sphene, and magnetite.

TABLE 3.—MINERAL COMPOSITION (MODE) OF THE QUARTZ MONZONITE PORPHYRY STOCK IN WEIGHT PER CENT

Specimen F collected 140 feet below a contact near the top of the stock; G collected from 230 feet below this contact. Both sites are about half a mile east of knob 9254. Precise locations are shown in Plate 1.

Mineral	F	G
Quartz	28.4	24.6
Plagioclase	42.4 (An 12)	33.9 (An 17)
Microcline	22.5	33.5
Muscovite	6.2	6.7
Biotite	0.5	0.8
Sphene	—	0.4
Apatite	t	t
Magnetite (?)	—	t
	100.0+	99.9+

The strong foliation suggests the intrusion of a viscous magma. Forceful upward movement is also indicated by the large block of Prospect Mountain quartzite along the northeast margin of the stock, for this block is at least several hundred feet above its normal stratigraphic position.

Composition and texture distinguish this body from the granodiorite stock to the north, though the composition of the quartz monzonite is similar to the border phase of the granodiorite. It is possible that they arise from a common source. Dating problems are alike for both stocks; a middle Tertiary date of emplacement best fits the regional igneous history.

Rhyolite (?) Dikes

Widely scattered northward-trending dikes, believed to be porphyritic rhyolite, occur in

the Minerva district; none are found in the granitic stocks. The dikes range from only a few feet to a few tens of feet in width and from a few hundred feet to possibly more than a mile in length.

TABLE 4.—CHEMICAL COMPOSITION CALCULATED FROM THE MODE OF THE QUARTZ MONZONITE PORPHYRY

Oxides	F	G
SiO_2	73.8	71.6
Al_2O_3	15.8	16.4
Fe_2O_3	t	0.1
FeO	0.1	0.1
MgO	0.1	0.1
CaO	0.1	1.2
Na_2O	4.4	3.3
K_2O	4.5	6.6
H_2O	0.3	0.3
TiO_2	t	0.2
P_2O_5	t	t
	100.0+	99.9+

Rocks of this dike group are strongly weathered and appear in various pale tints of brown, green, orange, and gray. Phenocrysts up to 3 mm in diameter, most of which are quartz and feldspar, constitute 5–10 per cent of the rock, and the groundmass contains feldspar microlites and vitrophyric interstitial material. The identification of this rock as a rhyolite remains tentative until more complete chemical data are available. Table 5 presents the mineralogy of the phenocrysts.

Three admittedly weak but independent lines of evidence suggest the association of the porphyritic rhyolite with granitic rocks of the stocks. The granitic sill in Big Wash appears to change its attitude and trends toward a rhyolite dike north of peak 11,064. The composition of these dikes is more similar to that of the granitic rocks than to that of the later flows. Furthermore, the absence of these dikes within the stocks may be explained by their contemporary or even slightly earlier intrusion.

Quartz Diorite Dikes

A few dikes with a moderately basic composition and a moderate to coarse pilotaxitic texture occur along the north fork of Lexington Creek and at other scattered localities. They are as much as a few tens of feet wide and also follow a northerly trend. Chilled borders are prominent, and the rocks appear greenish

TABLE 5.—ESTIMATED ABUNDANCE OF MINERALS PRESENT AS PHENOCRYSTS IN THE RHYOLITE PORPHYRY DIKES

Mineral	Volume Per Cent
Quartz	40–80
Plagioclase	20–40 (An 8-35)
Sanidine	0–25
Biotite, Muscovite	0–5

TABLE 6.—ESTIMATED MINERAL COMPOSITION AND ABUNDANCE OF THE QUARTZ DIORITE DIKES

Mineral	Volume Per Cent
Plagioclase	50–70
Hornblende	15–30
Quartz	5–15
Apatite, Sphene, Black opaques	0–5

gray. The amphibole and plagioclase laths contain interstitial quartz and additional feldspar. The plagioclase has an anorthite content of 25–35 per cent; the mineral composition of these dikes is summarized in Table 6. Several dikes along Lexington Creek contain quartzite inclusions that have greater textural affinities to the Prospect Mountain quartzite than to the Eureka quartzite. A finer-textured, more feldspathic phase of one of these dikes shows dilation relations toward the earlier, more normal phase. As these dikes cut the granitic bodies they are best dated as middle or late Tertiary.

FOLDS

Broadly speaking, the quaquaversal structure around the Wheeler Peak area recognized by Howell (1875) and the monocline at the southwest edge of the range mentioned by Spurr (1903) are accurate structural descriptions. From the Wheeler Peak area strata and thrust faults alike form a gentle, open anticline plunging southward about 5°. Dips

rarely exceed 25° except along the southwest edge of the range, where they steepen abruptly to 50° and 60°.

Superposed on this major structure are lesser eastward-trending folds. These have already been shown to postdate the thrust faults and probably are associated with the plutonic intrusion. The northward-trending fold north of Murphy Wash may be similarly explained, but no critical evidence is available yet.

Conglomerate of Late Tertiary(?) Age

Poorly sorted cobble conglomerate of undetermined thickness overlies the Paleozoic sedimentary rocks with major angular unconformity and is unconformably overlain by additional conglomerate and gravel. The older cobble conglomerate is well exposed in the hills north of the fish hatchery along lower Snake Creek. It includes all clastic rocks above the Paleozoic rocks and beneath the unconformity separating tilted from essentially flat-lying clastics.

The older conglomerate contains fragments of most of the rocks now exposed in this area. The conglomerate along the east flank of the southern Snake Range, however, contains limestone cobbles of Pennsylvanian age and cobbles of dacite porphyry. The nearest exposures of these rocks are 10 miles south and southwest in Johns Wash. A few rhyolitic tuff beds are associated with the conglomerate along the east flank of the range as at the dugway in Big Wash. Faults cut this conglomerate but do not extend into the overlying gravels. In the Johns Wash area a similar conglomerate underlies a dacite porphyry lava flow and contains cobbles of granite similar to that of the stock exposed south of Wheeler Peak.

Accurate dating of this conglomerate is not possible for no fossils have been recovered from it. At present a lake Tertiary(?) age seems the best choice. The conglomerate was probably subaerially deposited adjacent to a high range, perhaps during an earlier time when the Snake Range had a bold relief, for it closely resembles the more recent fan deposits west of the range.

Dacite Flow

The northernmost extension of the porphyritic lavas so widespread south of the Snake Range covers a portion of the Johns Wash—Murphy Wash area. There a flow overlies the conglomerate of late Tertiary(?) age and is tilted gently southward. The scattered cobbles of porphyritic dacite in the conglomerate on the east flank of the range indicate a former more extensive cover.

The flow in the Johns Wash—Murphy Wash area contains as much as 50 per cent phenocrysts that are partly fragmented and are embedded in a finely crystalline groundmass that has distinct flow structure. The euhedral quartz phenocrysts probably are the source of the quartz-crystal sands along the South Highline beach. Table 7 summarizes the composition of the phenocrysts.

Table 7.—Estimated Abundance of Minerals Forming Phenocrysts in the Dacite Porphyry Lava at the North and South Ends of the Flow in the Map Area

Mineral	North End Volume Per Cent	South End Volume Per Cent	Average Volume Per Cent
Plagioclase	55 (An 60)	65 An 27–30	60
Quartz	30	20	25
Hornblende	5	10	5–10
Biotite	5	5	5
Sphene	0–1	—	t
Black opaque	1–2	1–2	1–2

The source of the flow probably lies to the south where volcanic rocks are abundant. The position of the flow in a valley indicates that the topography at the time of extrusion was generally similar to that of the present. A late Tertiary or Pleistocene age is suggested for this flow.

Surficial Deposits

Gravel and partly consolidated conglomerate flank both sides of the range and occupy diverse elevations above the present stream channels. The high fan surfaces along Spring Valley clearly truncate the oldest beach of Spring Valley Lake, a now desiccated lake that probably was contemporary with Lake Bonneville of Wisconsin age (Hunt, Varnes and Thomas, 1953, p. 41). Most or all of the high-level gravels are probably of Pleistocene or Recent age.

Sand and gravel are being deposited on the youngest fans and along streams. Such deposits

grade into surficial mantle, talus, glacial till and outwash toward the range and into aeolian and lacustrine deposits toward the basins.

Beach and lacustrine deposits associated with the pluvial lakes in Spring Valley are exposed in the gravel pits that supply local road metal along the two superb beach bars, the North and South Highlines. The lenticular beds of sand, gravel, and small pebbles along the South Highline contain about 10 per cent of subhedral to euhedral detrital quartz that is probably derived from the dacite porphyry lavas in the hills to the south. Pelecypod fragments occur in the finer sand lenses. The North Highline consists of slightly coarser material, and some lenses contain pelecypods and gastropods. A clay-rich silt beneath the beach deposits may be of lacustrine origin.

Sand dunes cover a small fraction of Spring Valley. Longitudinal dunes with a trend of N. 15° E. grade into irregular lobes where sand is scarcer or vegetation more plentiful. These dunes and other partly anchored ones probably do not exceed 10 feet in height. The relative position of deflation centers or source areas of silt and fine sand with respect to the dunes indicates a prevailing northeasterly strong-wind direction at least since the desiccation of Shoshone Lake.

Erosional and depositional evidence of glaciation is restricted to those valleys near peaks at least 11,000 feet high. Cirques range in size from the large basin now filled with rock flow and talus debris northeast of Wheeler Peak to such small, dubious, or incipient basins as those southeast of Lincoln Peak.

Two sets of morainal ridges cross the upper reaches of the larger valleys. At and above about 9500 feet arcuate ridges of boulder till are separated by kettles that generally are internally drained. Moraine lakes are found in regions of this higher and younger moraine, here referred to as the Dead Lake Moraine.

At altitudes above 8100–8600 feet there are obscure remnants of the Lehman Creek Moraines. Though the topography is hummocky, drainage is well integrated. The till of these Moraines is also a boulder till but contains more abundant interstitial material and soil cover.

The Dead Lake and Lehman Creek Moraines are best tentatively correlated respectively with the Tioga and Tahoe Moraines in the Sierra Nevada on the basis of their topographic expression and similar topographic positions. If this correlation is valid, the Lehman Creek Moraines are possibly of early Wisconsin age, and the Dead Lake Moraines are possibly of late Wisconsin age.

In the Ruby Mountains 125 miles north-northwest Sharp (1938, p. 312) has a similar twofold division of Wisconsin moraines. Sharp estimates the lowest snow line at 9400 feet based on the summit altitudes behind the lowest cirques, and at 9800 feet based on cirque-floor altitudes. Comparable snow-line estimates for the southern Snake Range are 10,800 feet and 11,800 feet. The lowest recorded Wisconsin(?) snow line declines north-northwestward about 13.5 feet per mile between these ranges.

Normal Faults

Normal faults are commonly associated with the borders of the southern Snake Range and generally have a northerly strike. Major movement along the west flank of the range occurred on a fault or faults now completely buried. The steepness of the scarp and the alignment of the toes of the spurs across rocks of various lithologies suggest that a high-angle fault lies close to these spurs. Presumably this fault branches or bends near the northern part of the area, for the range front shifts several miles westward north of the map area. The displacement of this high-angle fault is unknown, but the 7000 feet of relief may be a minimum value. Several high-angle normal faults parallel this border fault, and a set of low-angle normal faults that show considerable strike-slip displacement are prominent in the Minerva district.

On the east flank of the range there is no single large straight scarp; instead there are several echelon scarps that are paralleled by branching faults. South of Lexington Creek the Fish Haven dolomite on the east side is downthrown against the Lincoln Peak formation. The stratigraphic displacement there may be 1000–3000 feet depending on the stratigraphic omission caused by thrust faults. Along Big Wash the conglomerate of late Tertiary(?) age is faulted against limestone breccia of the Pogonip group, and again the amount of displacement, though uncertain, must be large. The graben complex along Cave Canyon is probably a continuation and bifurcation of one of these border faults. The northward extension of this graben complex cuts or is cut by a tongue of the granodiorite; the full extent and origin of the shattering of this tongue is unknown. About 1½ miles east of

the fish hatchery another high-angle fault is indicated by the alignment of outlying knobs. The topographic asymmetry of the range suggests that eastward or southeastward tilting of the range accompanied the normal faulting.

Precise dating of the various normal faults along the borders of the range is not possible for the record of young deposits is incomplete and also without precise dates. Several episodes of normal faulting can be demonstrated along these borders: by the three sets of mutually unconformable gravels and young conglomerates and by the several erosion surfaces east of the range. The oldest of these episodes may be earlier than the granodiorite stock; the later ones certainly are much younger than this stock. At best these normal faults may be fitted into the regional picture and tentatively dated as middle and late Tertiary and Quaternary.

Historical Summary

In the southern Snake Range the record of geologic events begins with an almost uninterrupted sequence of sedimentation. In conjunction with regional evidence, the Prospect Mountain quartzite of Cambrian age probably marks the initial eastward transgression of a shallow sea that later deposited lime and minor amounts of mud. During Middle Ordovician time the Eureka quartzite was probably deposited by a regressive sea, and local emergence and a little erosion may have interrupted the depositional sequence. Dolomite first appears in Lake Ordovician time and continues into the Devonian, after which limestone again is more common. The increasing importance of clastics in Carboniferous time reflects a rising land mass to the west. A short distance south of the map area this depositional sequence continues into the Permian.

Between Late Cambrian and the time of thrust-fault movement the base of the stratigraphic sequence was slightly metamorphosed.

A major orogenic episode followed. In Mesozoic or early Tertiary time, one period of east-west compression produced complex thrust faults. The planes of fault movement shifted progressively to higher stratigraphic levels. A near-surface environment might account for the extensive brecciation and normal faults that accompany the thrust faults. Slight folding along north-south axes accompanied or followed this period of compression.

Probably in middle Tertiary time a granodiorite and a quartz monzonite stock were intruded at low temperature. Their forceful injection resulted in the gentle warping of the overlying strata and thrust planes, and stoping removed much of the host rocks. The entire area was gently domed at this time. Rhyolite porphyry dikes may have accompanied the intrusion of the granodiorite and quartz diorite dikes postdate these stocks.

A second episode of fault movement probably began in Middle or late Tertiary time. Stress was relieved along high-angle northward-trending faults, and the range was raised and tilted gently southeastward. Successive movements in response to this stress are recorded by coarse conglomerate and gravel, by a large angular unconformity within the conglomerate and gravel by normal faults of various ages cutting these clastics, and probably by the deep incision of the young conglomerate and gravels. A dacite porphyry lava flow and associated tuff is contemporaneous with the oldest of these clastics.

During the Pleistocene epoch the higher valleys contained glaciers that mark two periods of ice advance. Pluvial lakes in Spring Valley probably accompanied these glacial maxima. The return to a drier, warmer climate led to the desiccation of the lakes and the disappearance of the ice. Sand dunes and alluvium are now being deposited in the lake basin.

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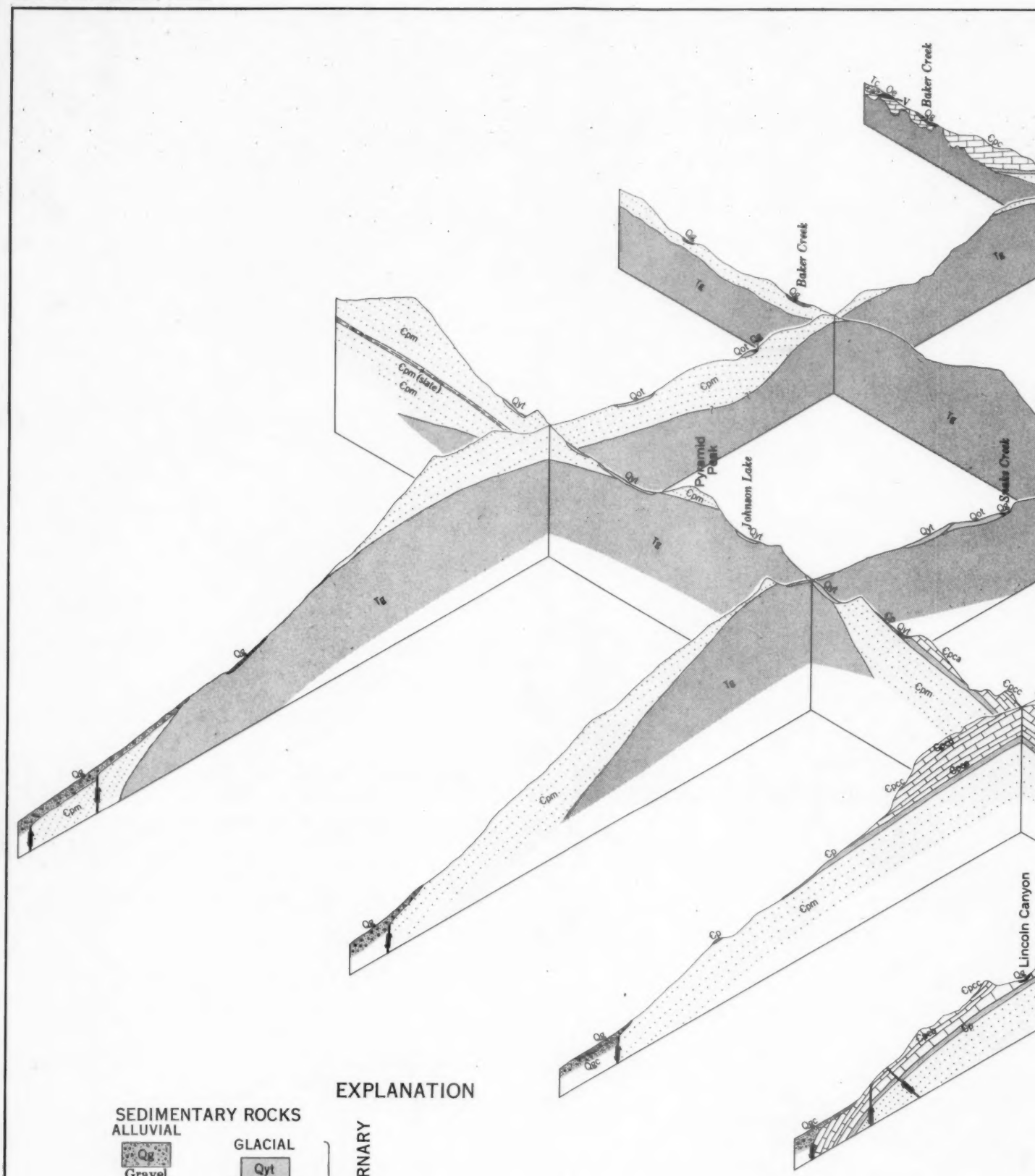


EXPLANATION

SEDIMENTARY ROCKS

ALLUVIAL

- Qg — Gravel
- Qgc — High level gravel and conglomerate

GLACIAL

- Qyt — Younger till
- Qot — Older till

QUATERNARY

ANGULAR UNCONFORMITY

- Tc — Conglomerate

TERTIARY(?)

ANGULAR UNCONFORMITY

- Cs — Shale, limestone and sandstone
- Guilmette formation(?)
- Fish Haven, Laketown, Sevy and Simonson dolomites undifferentiated

DEVONIAN MISS. AND TO DEVONIAN PENN.(?)

POSSIBLE DISCONFORMITY

- Eureka quartzite
- Op — Pogonip group
- Ccs — Corset Spring shale
- Clw — Johns Wash limestone
- Clp — Lincoln Peak formation

ORDOVICIAN

CAMBRIAN

Middle Cambrian

- Cpce, Cpcd, Cpc, Cpcc, Cpcb, Cpca — Pole Canyon limestone
  Members a through e distinguished locally.

Lower Cambrian

- Cp — Pioche shale
- Cpm — Prospect Mountain quartzite

Upper Cambrian — *Lower Ordovician* — *Middle Ordovician*

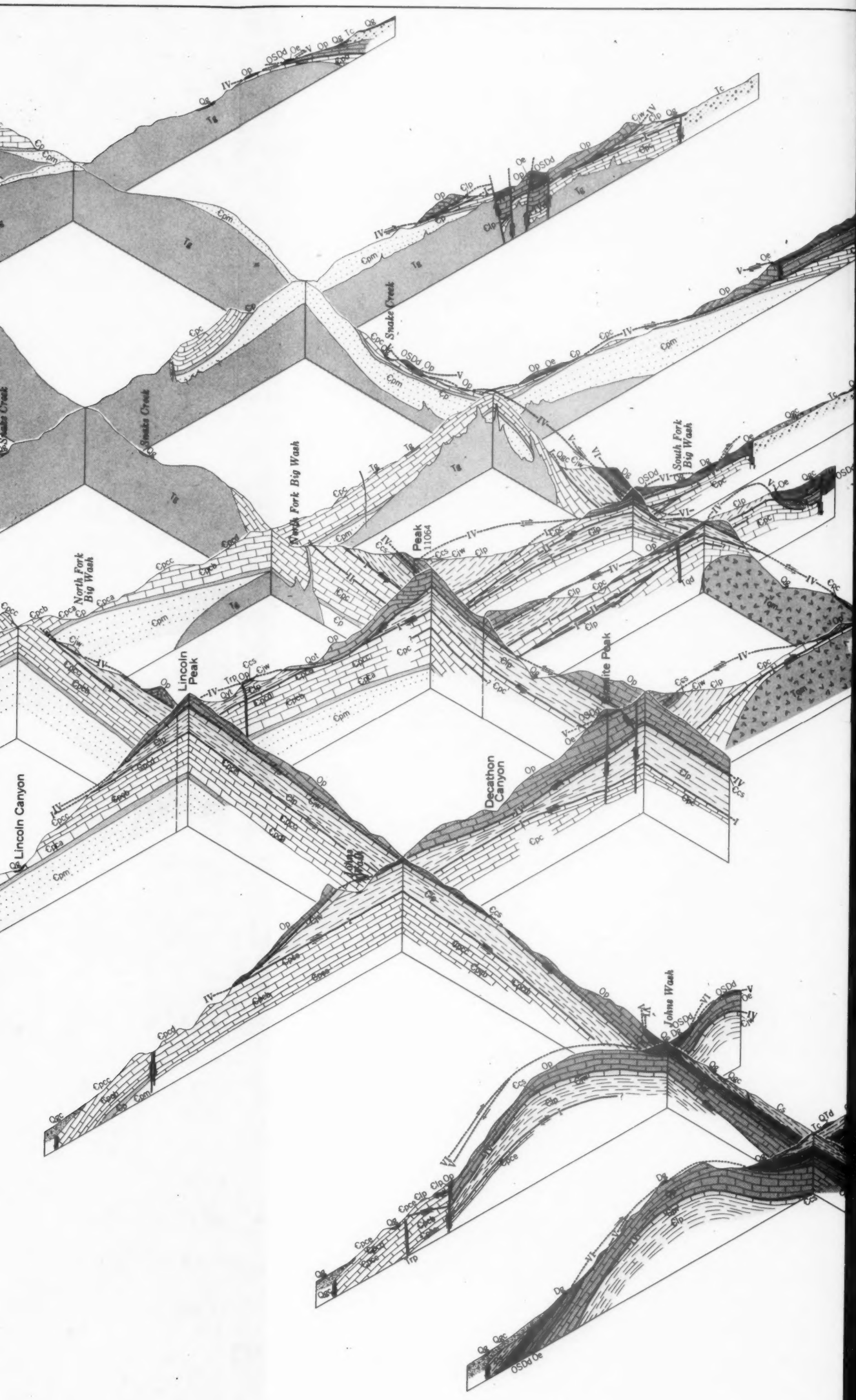
CAMBRIAN

IGNEOUS ROCKS

- Tqd — Quartz diorite dike
- Trp — Rhyolite porphyry dike
- Dacite porphyry flow
- Tg — Granodiorite stock
- Tqm — Quartz monzonite porphyry stock

TERTIARY(?) TERTIARY OR QUATERNARY

FENCE DIAGRAM



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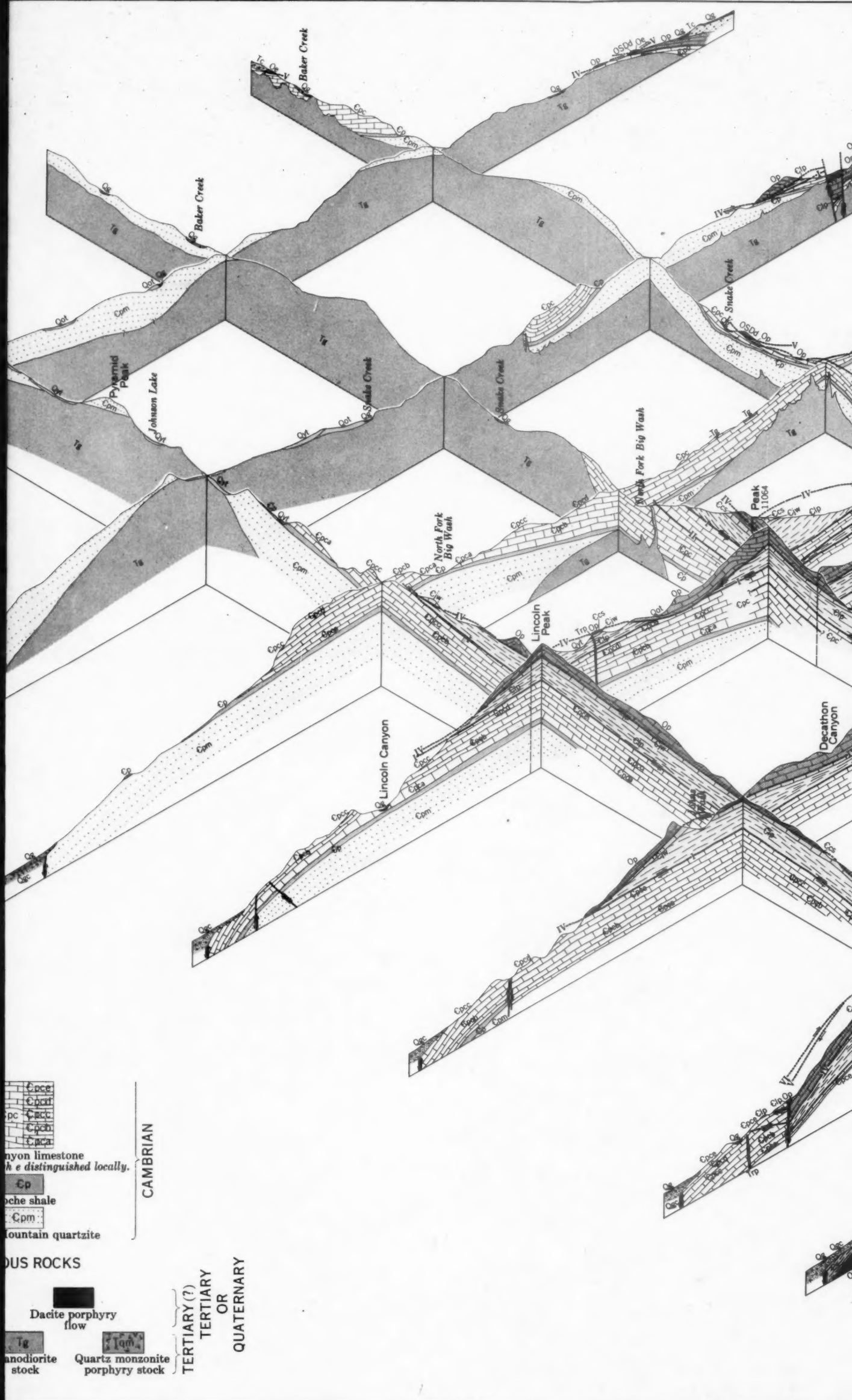


nyon limestone
h e distinguished locally.

Cp
oche shale

Cpm
lountain quartzite

CAMBRIAN

OUS ROCKS

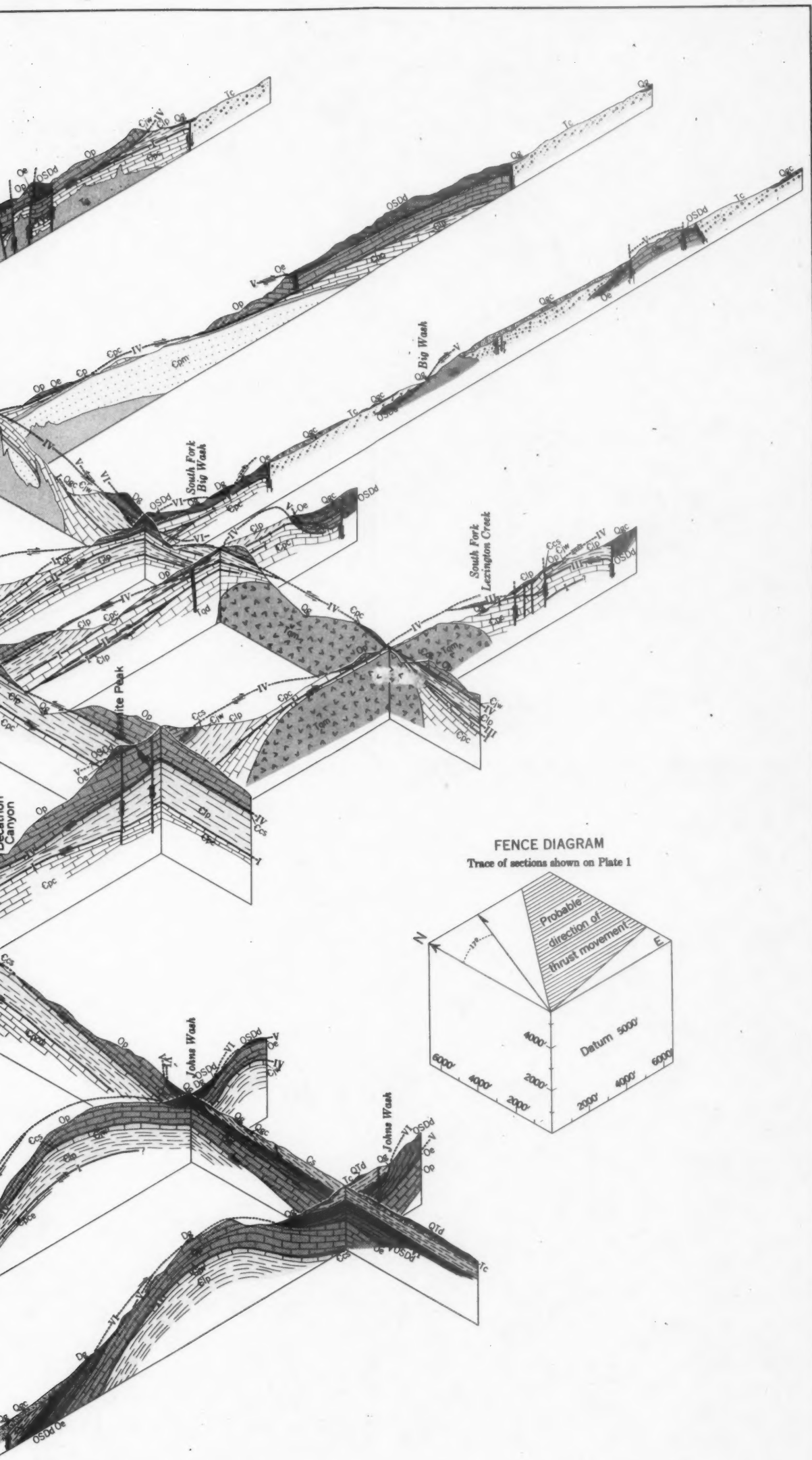
Dacite porphyry flow

Tg
anodiorite stock

Tqm
Quartz monzonite porphyry stock

TERTIARY(?)
TERTIARY OR QUATERNARY

FENCE DIAGRAM OF THE SOUTHERN SNAKE RANGE, NEVA



FENCE DIAGRAM

Trace of sections shown on Plate 1

EVADA